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AFCRL—63—435

MASSACHUSETTS INSTITUTE OF TECHNOLOGY DEPARTMENT OF METEOROLOGY

CONTRACT NO. AF19(604)-5223
PROJECT 8628 TASK 862807

FINAL REPORT
PLANETARY CIRCULATIONS PROJECT

VICTOR P. STARR, DIRECTOR

30 NOVEMBER 1962

Prepared for

THE GEOPHYSICS RESEARCH DIRECTORATE
AIR FORCE CAMBRIDGE RESEARCH LABORATORIES
OFFICE OF AEROSPACE RESEARCH
UNITED STATES AIR FORCE
BEDFORD, MASSACHUSETTS

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STUDIES OF THE STRATOSPHERIC GENERAL CIRCULATION

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ABSTRACT

The aim of the studies reported under this contract was to elucidate the various mechanisms by which the budgets of energy, angular momentum and mass for the stratospheric region are satisfied. In this respect the objectives coincided with the purpose of work done simultaneously under AEC support. Similar studies for the atmospheric regions from the surface up to 100 mb have been reported in detail in previous Final Reports (Studies of the Atmospheric General Circulation I, II and III, 1954, 1957, and 1959 under contracts AF19(122)-153, AF19(604)-1000, and AF19(604)2242) and it will be recalled that a general finding was that large-scale quasi-horizontal eddy processes proved to be the principal agents in the transfer of angular momentum, in the generation of mean zonal kinetic energy and in the transfer of heat energy. Prior to the International Geophysical Year many of the workers who discussed the stratospheric general circulation invoked mean meridional motions in an essentially quiescent region to satisfy what were thought to be the requirements for transfer. Thus there appeared to be a basic difference in the predominating motion systems in the two regions and the cause was generally ascribed to the large static stability of the stratosphere as compared with the troposphere. The object in the present studies was to make diagnostic calculations of terms in the energy and momentum budgets of the stratosphere using the techniques previously applied to the troposphere

and with as large a data sample as possible. The International Geophysical Year was selected for study because a special effort was made throughout that time to attain high altitudes with the wind and temperature sensors and all data were carefully checked and published by the World Meteorological Organization.

The report opens with an examination by Professor Starr of the possible role of the vertical advection of kinetic energy in the stratospheric energy budget; the paper raised questions that we sought to answer from the observations reported in the following papers.

The second paper, by Barnes, is a detailed study of the energy budget of the northern hemisphere stratosphere during the first six months of the IGY. Barnes' paper includes a description of the data-processing techniques used in all of the studies herein. All the terms in the budget which can be evaluated from wind and temperature observations are considered. Terms including diabatic heating could not of course be evaluated on a day-to-day basis but for some of them estimates have been included which apply to long periods. It appears from Barnes' work that both Reynolds' stresses and Coriolis forces acting on mean meridional motions are of importance in the energy budget. Many of the terms in the energy budget are estimated from data for the first time; in this respect it should be noted that diabatic effects are thought to be smaller in the lower stratosphere than elsewhere in the atmosphere but nevertheless as Barnes shows they cannot be ignored.

The third paper, by Dickinson, examines the stratospheric angular momentum budget for the first year of the IGY period. In this case quasi-horizontal eddies seem to be of paramount importance and the mean meridional motions necessary to accomplish a balanced budget depend very strongly on the values selected for the vertical eddy diffusion coefficient. If these are assumed to be large ($\sim 10^6 cm^2 sec^{-1}$) then significant mean meridional motions are involved but for small values ($\sim 10^3 cm^2 sec^{-1}$) the mean motions are extremely small ($<$ several cm sec^{-1}).

The fourth paper, by Oort, examines the mean meridional motions obtained directly from the wind observations. There appears to be a continuation upwards of the three-cell structure found in the troposphere with equatorward flow throughout the year in middle latitudes. The finding is contrary to the early speculation, mentioned previously, concerning the direction of mean meridional motions in the stratosphere. The equatorward mean motions in middle latitudes are about 30 cm sec^{-1} and would imply correspondingly large coefficients of vertical eddy transfer in Dickinson's analysis. There is a possibility that the observations are biased by the presence of a zero wave number tidal component which is in the same direction at both times used (00 and 12Z). This point is being pursued further; it could be resolved if there were sufficient stations reporting at 06 and 18Z but such is not the case.

The fifth and sixth papers, by Newell, are concerned with trace substances. They illustrate an application of the stratospheric data and provide some indirect information on the mass budget of the stratosphere. It is of considerable importance to determine the way in which stratospheric motions influence the distribution of trace substances such as ozone and water vapour as these distributions play a large part in the radiative heat budget of the region. An approximate ozone budget is constructed whose main component involves the transfer by quasi-horizontal large-scale eddies. The manner in which these eddies can produce a spring maximum in both ozone and fission-product radioactivity, as well as a middle-latitude maximum in the latter, is also discussed. Trace s ubstance information appears to support low values of the diffusion coefficient discussed above and hence indirectly is further evidence in favour of small-magnitude mean meridional motions.

The papers in the appendix while not presenting climatological results are nevertheless pertinent to the information discussed herein. Barnes' paper is an out-growth of his analysis work with the IGY data and discusses a topic which is normally avoided, namely the meaning and treatment of the components of the wind vector u and V in the vicinity of the singularity points at the earths' poles. White and Nolan's paper is a case study of the energy conversion process in the stratosphere. It compliments the climatological information presented by Barnes.

The papers included herein treat particular aspects of the data. It is planned to publish the processed data in toto in a series of three atlases, each covering a six-month portion of the IGY. The first of these, authored by Dr. Takio Murakami, appeared in March 1962 and the others are in preparation under AEC sponsorship. This basic information will be of value for a number of further studies; in fact we have only scratched the surface as far as the interpretation of the results is concerned.

It is possible to make comparisons between these data and the hydrodynamic model experiments and one example is discussed in Oort's paper. The diagnostic results can also be used in numerical models. Both of these approaches can help towards the goal of an ultimate understanding of the physical reasons for the observed circulation. The data may also be of use in interpreting studies of the circulation of other planetary atmospheres where observations are only available from the higher levels.

Thus although the present report completes our formal submission of work done under the contract we shall continue to use the data as a basis for a variety of further studies (many of which are already in progress under AEC sponsorship) giving due acknowledgement to the contract where appropriate. One specific example is a further study of the ozone question.

Reginald E. Newell

Nicht im Handel

Reprint from

Archiv für Meteorologie, Geophysik und Bioklimatologie

Serie A: Meteorologie und Geophysik, Band 12, 1. Heft, 1960

Herausgegeben von

Doz. Dr. W. Mörikofer, Davos, und Prof. Dr. F. Steinhauser, Wien

Springer-Verlag in Wien

551.511.3:551.510.53

(Massachusetts Institute of Technology, Cambridge, Mass., U.S.A.)

Questions Concerning the Energy of Stratospheric Motions[1]

By

V. P. Starr

Summary. In this article the theoretical conclusion is reached that the vertical transport of kinetic energy in the atmosphere from one horizontal layer to another is effected solely by an area integral over the internal boundary, of the kinetic energy per unit volume multiplied by the vertical velocity. Speculations are made as to whether the kinetic energy in stratospheric levels is maintained against friction through such a vertical transport from other levels, or whether the needed supply is derived from convective sources *in situ*.

Zusammenfassung. Auf Grund theoretischer Überlegungen wird der Schluß gezogen, daß der Vertikaltransport kinetischer Energie in der Atmosphäre von einer horizontalen Schicht zu einer anderen einzig durch ein Flächenintegral über die Grenzfläche bestimmt wird, dessen Integrand das Produkt aus kinetischer Energie pro Volumeinheit und der Vertikalgeschwindigkeit ist. Es werden Überlegungen angestellt, ob die kinetische Energie in Stratosphärenschichten gegenüber der Reibung durch einen solchen Vertikaltransport aufrechterhalten werden kann oder ob der benötigte Nachschub von konvektiven Quellen in der Schicht selbst stammt.

Résumé. Se fondant sur des considérations théoriques, l'auteur arrive à la conclusion que le transport vertical d'énergie cinétique dans l'atmosphère, d'une couche horizontale à une autre, est uniquement déterminée par une intégrale de surface étendue à la surface limite et dont l'intégrande est égale au produit de l'énergie cinétique par unité de volume par la vitesse verticale. Il discute la question de savoir si l'énergie cinétique d'une couche stratosphérique, en raison du frottement, est maintenue par un tel transport vertical ou si l'énergie d'entretien provient de sources convectives de la couche elle-même.

[1] The research reported in this paper has been made possible through the support of the Geophysics Research Directorate of the U. S. Air Force under contract AF 19(604)-5223.

1. Introduction

Meteorologists, when writing about the phenomena of the stratosphere are often wont to speculate concerning such things as the energy sources for the motions in the higher regions of the atmosphere. These ponderings are usually, and perhaps understandably, rather vague and diffuse—what with the lack of proper observational information and of a suitable theoretical framework within which observations might give expression to significant over-all physical facts. But more and more soundings reaching high altitudes are now being made on a routine basis, and it seems to me that the formulation of precise questions, grounded in the most reliable principles of physics, is the primary business of the research worker in this field who wishes eventually to secure some degree of rational understanding. The principles of physics might be such statements as those of the conservation of mass, of momentum and of energy, etc. The problem requiring resourcefulness is to formulate them in such a fashion as to make possible the observational study of the manner of their fulfilment, and to arrive thereby at some nontrivial physical conclusions pertaining to the real atmosphere. The eventual accumulation of such deductions should then lead, ideally at least, to the synthesis of quantitative models of various types (see, e. g., STARR [2]). What follows is an attempt to take one simple step along such a road.

2. Physical and Mathematical Considerations

We shall begin by considering the entire atmosphere as being divided into two horizontal layers by a closed constant level surface located at some appropriate fixed geodynamic height above sea level. Assuming that this constant elevation is plausibly chosen, we may for the purpose of our discussion name the entire upper region the "stratosphere," and the lower one the "troposphere." We realize, of course, that certain liberties are thus taken with standard terminology, although it may later prove that something less drastic is also amenable to precise treatment.

A number of years ago the writer had occasion to discuss the mechanical energy equation for the horizontal components of motion in the atmosphere (STARR [1]). Let us write this equation for the stratosphere in the form

$$\frac{d}{dt}\int \rho \frac{u^2+v^2}{2}\, d\tau = \int \rho \frac{u^2+v^2}{2}\, V_n\, ds + \int p \operatorname{div} \vec{V}_h\, d\tau - D \qquad (1)$$

Here u, v are the eastward and northward components of particle velocity, ρ is density, p pressure, t time, $d\tau$ a volume element, ds an element of surface of the volume taken, V_n the inward normal component of particle velocity across the boundary surface, $\vec{V}_h$ is the horizontal wind vector and D is the total rate of frictional dissipation of kinetic energy in the volume. The genesis of eq. (1) from the equations of motion for the horizontal directions and the general continuity equation is rather obvious

and will not be repeated here. It contains the statement for our problem that the amount of kinetic energy of horizontal motions in the stratosphere may change (left-hand side) as a result of advection of such energy across the boundaries of the stratosphere (first term on right), or as a result of sources of kinetic energy in the volume of the stratosphere (second term), or as a result of work done against friction (third term).

In the long term average the total kinetic energy in the stratosphere neither increases nor decreases, except for possible minute effects which reflect a changing climatic regime. On the other hand the dissipation must continuously be degrading the energy of stratospheric motions. Hence if A denotes the advection of kinetic energy into the stratosphere and S denotes the source term we may write that for the long term mean

$$D = A + S \tag{2}$$

that is, *the dissipation must be made good either by an advection effect or by a generation of kinetic energy in situ.* Our entire conception of the operation of the stratospheric circulation depends crucially upon the elucidation of how this equation or some equivalent of it is satisfied in actuality.

Let us examine first the advective effect A for the stratosphere. The upper surface may be eliminated because only negligible amounts of kinetic energy are exchanged with outer space (meteors and the like). This leaves only the horizontal internal boundary to be reckoned with. It follows therefore that A may be written as follows,

$$A = \int \frac{u^2 + v^2}{2} \rho\, w\, d\sigma \tag{3}$$

where $d\sigma$ is an element of area of the internal boundary and w is the upward component of particle velocity. In the average we may, with a high degree of precision, write that

$$\int \rho w\, d\sigma = 0 \tag{4}$$

from continuity considerations. Comparison of (4) with (3) now leads us to the purely mathematical conclusion that if A is to be non-zero, then there must exist a correlation between $(u^2 + v^2)/2$ on the one hand and ρw on the other. Thus if A were to be positive, ρw would have to be positive by and large over those regions where $(u^2 + v^2)/2$ is relatively large.

It is to be specifically stressed at this point that, due to the physical circumstances at *any* constant level in the atmosphere, the exchange of kinetic energy in the vertical takes place through the simple action specified by eq. (3), without any other significant *transport* mechanism.

The second term on the right side of eq. (1) which was designated as S in eq. (2) has been discussed previously by the writer in the reference given and also in other connections. It is a volume integral which may be written with sufficient accuracy for our problem in the form

$$S = \int [\int p \operatorname{div} V_h \, d\sigma] \, dz \tag{5}$$

in which $d\sigma$ is again an element of horizontal area and dz is an element of vertical distance. Since we know that

$$\int \operatorname{div} V_h \, d\sigma \equiv 0 \tag{6}$$

we may reason about the eqs. (5) and (6) in the same manner as we did in regard to (3) and (4). We see that in order for S to have a non-zero value the pressure and the divergence should, at least at some level be correlated in a spacewise sense.

In our problem the value of the expression (5) for S is difficult to estimate from general information or from data, unless perhaps its form is first changed in order to eliminate the necessity of measuring the horizontal divergence. Let us try to do this. We may first write that

$$\int p \operatorname{div} \vec{V}_h \, d\sigma = - \int V_h \cdot \Delta_h p \, d\sigma \tag{7}$$

since the terms formally representing the difference between these expressions integrate to zero for our region. We now define

$$\omega \equiv \frac{dp}{dt} = \frac{\partial p}{\partial t} + \vec{V}_h \cdot \nabla_h p + w \frac{\partial p}{\partial z} \tag{8}$$

whereupon it is seen that

$$S = - \int \omega \, d\tau + \frac{d}{dt} \int p \, d\tau + \int w \frac{\partial p}{\partial z} d\tau. \tag{9}$$

The second integral on the right vanishes since the long term average of the pressure is nearly constant. So also does the third integral which, on hydrostatic principles, is nothing more than the rate of change of the total potential energy of the air in the region (or essentially its total vertical momentum multiplied by the acceleration of gravity).

Only the first term now remains. By use of the thermodynamic coordinate p in the place of the geometrical vertical coordinate z, hydrostatically, we have

$$S = - \int \omega \, d\tau = - \int_0^{p_0} \int \omega \alpha \, d\sigma \, dp. \tag{10}$$

In (10) α is the specific volume, p_0 is the pressure at the internal boundary and $d\sigma$ here again stands for an element of horizontal area but now following a given pressure surface. The last term is the same quantity as has been used, for example, by White and Saltzman [*5*] to measure the generation of kinetic energy in the troposphere.

For a given closed isobaric surface in the stratosphere it may be shown that

$$\int \omega \, d\sigma = 0. \tag{11}$$

This follows directly through an integration of the continuity equation in the form

$$\frac{\partial w}{\partial p} + \nabla \cdot \underset{\to}{V_{h'}} = 0 \tag{12}$$

where the divergence of the wind is measured along an isobaric surface. From considerations which have recurred previously in this discussion, a comparison of the inner integral of the last term in (10) with eq. (11) shows that a given closed isobaric surface in the stratosphere can make a non-zero contribution to S, if and only if there exists a spacewise correlation of ω and α over the area of the isobaric surface. The fact that a few isobaric surfaces are not closed, but intersect the base of the stratosphere probably does not interfere much with this mode of thinking about the significance of the $\omega\,\alpha$ integral.

We observe that a correlation of α with ω as here discussed is a convenient means for the specification of a basic convective process. Thus if the negative values of ω are predominantly associated with large values of α, there exists a positive generation of kinetic energy. This presumably would signify a preponderance of rising motions of warmer air and a corresponding sinking of equal masses of colder air. It appears that for this form of S there is present some chance of a successful observational assessment of the energy producing action in the stratosphere.

3. Discussion of Conditions in the Real Atmosphere

I must confess that at least at present adequate stratospheric measurements of either A or of S are not at my disposal. Some small beginnings in this direction are, however, being made. WHITE and NOLAN [4] have been interested in measuring S, while certain other workers have, at my suggestion, become interested in appraising the observational possibilities of measuring A.

In a general sense the classical view has been that at least the lower stratosphere is a passive region wherein any tendency toward direct convective action is suppressed by the large hydrostatic stability present. If this is a dominant characteristic, then S should be zero or negative making the portion of the atmosphere above, let us say, 16 km a region of forced motion *on the average*. Such a view is strengthened by the circumstance that there exists a countergradient northward flow of heat at these levels (see, e. g., WHITE [4]).

If the notion of an inert, passive stratosphere corresponds to real fact, then we are at once brought to the concept that the continuance of motions in that region is to be explained by a vertical transport from other convectively more active layers above or below, through the action of a process represented by an integral like A. In that case although it cannot be gainsaid that still higher layers might contain sources of kinetic energy which could be transported downward, still the first consideration might be given to the hypothesis that the needed supply originates in the troposphere and is fed upward across levels such as 16 km or thereabouts.

4. Some General Comments

Due to the pivotal nature of the matters dealt with in the preceding discussions, several additional sidelights are not without justification. The following ones present themselves to the writer.

(a) Although our data still leave much to be desired, the time is coming shortly, if indeed it is not already here, when the evaluation of processes such as those dealt with in this paper should be given a high priority in meteorological research endeavors. Even though finally conclusive results may not appear immediately from such efforts, the partial insights gained would still be of no small importance, and should encourage thinking of an adequate scope and suitable perspective. We have seen somewhat comparable sequences of events during the past decade or two concerning other matters pertaining to the mechanics of the general circulation.

(b) In the manipulations the frictional term D is simply the work done by the fluid against frictional stresses arising from the viscosity of the fluid. No particular mathematical form for this action is assumed here, nor is any assumption made concerning the disposal of the energy involved—it may either remain in the fluid in the form of heat, potential energy etc., or it may be communicated to contiguous fluid masses by the frictional action itself. An example of the latter effect is the frictional transmission of kinetic energy from the lower atmosphere to the oceans which therefore *gain* energy thereby. This can of course happen only because the sea surface moves in response to the surface stresses which can therefore perform work upon the water.

It is an open question whether a similar action might not take place at the top of the troposphere, so that the stratosphere is dragged around by friction in the same manner as the oceans. Much here depends on what we might conceive as comprizing friction. If we limit the term to mean only molecular viscosity, then the drag due to it is no doubt much too small to cause concern and all other actions would be included in A. More usually meteorologists include as friction all rather small scale eddy effects such as those found in the so-called friction layer near the ground. With such a convention the frictional drag across levels such as 16 km would probably still be quite small, on the average. Any action from eddies of appreciable size would again be included in A.

The interpretation of the entire quantity A as a sort of gross friction is not very helpful, if for no other reason simply because we do not know even the sign of the viscosity coefficient which would be involved. Besides, we desire to know the details of the vertical velocity distribution and of its correlation with the kinetic energy.

(c) The primary convective actions in the troposphere take place at appreciable elevations above the earth's surface. In any event they do not take place in their entirety within the confines of the friction layer. Yet it is true that a disproportionately large fraction of the total dissipation of kinetic energy does take place in this bottom layer. It there-

fore follows that an integral of the form A acting across moderate elevations within and above the friction layer must represent an extremely important link in the total workings of the general circulation. The advisability of the detailed study of this anticipated phenomenon is manifest. Plans are being made currently to investigate it from data.

(d) It may be noted that the material treated is primarily of a mechanical nature. No equation of state has been used, and hence the conclusions are not dependent upon any such particular equation for the atmosphere which be assumed.

(e) Eq. (1) requires no assumption in regard to the presence or absence of hydrostatic balance, although the further manipulations of S do depend upon this condition being present. Likewise the validity of eq. (1) is in no way contingent upon the smallness of the vertical velocities in the atmosphere, although this latter is generally the case.

(f) As was stated in the discussion of 1948 by the writer, transports of kinetic energy horizontally across, let us say, vertical boundaries within the atmosphere may be effected through the work done by pressure forces in virtue of normal velocity components. Such a term arises in addition to the advective term in such a case.

Such a pressure-work term is absent in our equations for the transport of kinetic energy across horizontal surfaces. A pertinent consideration in this connection is that in our present case the pressure-work term would be capable of transferring kinetic energy of vertical motion only. Even for this certain departures from hydrostatic balance would enter. This explains its absence from our discussions in this paper which deals with the kinetic energy of horizontal motions only.

(g) It must be realized, however, that when the foregoing analysis is performed in pressure coordinates instead of geometric ones, the simplicity and directness of the results are to some degree lost. Thus an additional boundary term involving the product ω and the geopotential appears. In view of this fact one may question the adequacy of the approximation made when a constant pressure surface is arbitrarily substituted for the constant level bottom boundary.

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Printed in Austria

Kinetic and Potential Energy Between 100 mb and 10 mb

During the First Six Months of the IGY [(1)] [(2)]

by

Arnold Appleton Barnes, Jr.

ABSTRACT

Using the x,y,p coordinate system, the integral equations of the time rate of change of (1) total kinetic energy, (2) potential plus internal energy and (3) zonal kinetic energy for the stratosphere over the northern hemisphere are given in a form suitable for numerical approximation of the individual integrals. The method of approximation separates each integral into a mean term contribution and contributions due to the eddy terms. Observations at 100,50, 30 and 10 millibars from 211 northern hemisphere stations for the periods July through September 1957 and October through December 1957 were used to evaluate the terms.

The large hemispheric increase of total kinetic energy of the stratosphere observed from October through December 1957 was not due to vertical advection of kinetic energy but rather to conversion of potential plus internal energy into kinetic energy.

(1) The material in this article appeared as part of the author's doctoral thesis (Barnes, 1962).

(2) Present affiliation: Meteorological Research Laboratory, Geophysics Research Directorate, Air Force Cambridge Research Laboratories, Bedford, Mass.

For the period July through December 1957 in the stratosphere, the eddy kinetic energy did not provide sufficient energy to maintain the zonal kinetic energy. Evidence is presented indicating that the meridional circulation, working through the Coriolis transformation terms, was responsible for converting potential to zonal kinetic energy to maintain the observed level of zonal kinetic energy.

Diabatic motions in the lower stratosphere are as important as the adiabatic vertical motions when considering the mean meridional circulation. Temporal variations of the net radiation are very important in determining the diabatic motions.

Kinetic and Potential Energy Between 100mb and 10mb

During the First Six Months of the IGY

I. Introduction.

The purpose of this paper is to give an empirical description of the mechanical energy balance of the stratosphere over the northern hemisphere for the first six months of the International Geophysical Year (IGY), viz. July 1957 through December 1957.

One of the most successful ways of studying the atmosphere has been the statistical reduction of the observed upper-air data over a large area and a long time period. V.P. Starr and his colleagues have used this technique for the study of the water vapor balance, the energy balance and the angular momentum balance of the atmosphere over the northern hemisphere.

Because of the lack of data, these studies, until very recently, have been confined to the troposphere. The special effort put forth by the countries participating in the IGY upper-air program produced enough data over the northern hemisphere to allow us to investigate the stratosphere by those same methods which have been so successful in the lower reaches of our atmosphere. It is felt that the IGY coverage and amount of data at and above 100 mb is the best that will be available for some time to come. Special efforts were made to obtain stratospheric data from areas of sparse coverage, but, because of the philosophy behind such studies, extrapolation and interpolation of data were not used except in the analysis of various maps in the final stages of the study.

For the study of the atmosphere, certain integral equations were derived (see Chapter II). These integrals were then evaluated by

finite difference methods using the raw data.

Using the x, y, p coordinate system, the time rate of charge of the kinetic energy of the atmosphere north of a latitude ϕ_0 and between two pressure layers p_L and p_u may be expressed as

$$\frac{\partial}{\partial t}\iiint E\frac{dp}{g}\,dx\,dy = \iint E V\frac{dp}{g}\,dx - \iint\left(\frac{E\omega}{g}\right)_L dx\,dy \tag{1.1}$$
$$+ \iint\left(\frac{E\omega}{g}\right)_u dx\,dy + \iint \psi V\frac{dp}{g}\,dx - \iint\frac{(\psi\omega)_L}{g}\,dx\,dy$$
$$+ \iint\frac{(\psi\omega)_u}{g}\,dx\,dy - \iiint\frac{\omega\alpha}{g}\,dp\,dx\,dy - \iiint\frac{\mathbb{V}_p\cdot\mathbb{F}}{g}\,dp\,dx\,dy$$

where $E \equiv \frac{\mathbb{V}_p\cdot\mathbb{V}_p}{2}$. This equation is derived and discussed in detail in Chapter II.

Similarly, the time rate of change of the sum of the potential and internal energy of the same region may be expressed as

$$\frac{\partial}{\partial t}\iiint(\mathcal{P}+\mathcal{J})\frac{dp}{g}\,dx\,dy = \iint\frac{c_pTV}{g}\,dx\,dp - \iint\left(\frac{c_pT\omega}{g}\right)_L dx\,dy \tag{1.2}$$
$$+ \iint\left(\frac{c_pT\omega}{g}\right)_u dx\,dy + \frac{p_L}{g}\iint\left(\frac{\partial\psi}{\partial t}\right)_L dx\,dy - \frac{p_u}{g}\iint\left(\frac{\partial\psi}{\partial t}\right)_u dx\,dy$$
$$+ \iiint\frac{\omega\alpha}{g}\,dp\,dx\,dy + \iint\frac{dQ}{dt}\frac{dp}{g}\,dx\,dy.$$

By finite difference methods, we evaluated the time averages of all of the double integrals of equations (1.1) and (1.2). The method used allowed us to separate the contributions of some of the integrals into mean terms, standing eddies and transient eddies.

The triple integral

$$\iiint \frac{\omega \alpha}{g} \, dp \, dx \, dy \tag{1.3}$$

which appears in both of the above equations was also evaluated. This term represents a conversion between kinetic energy and potential plus internal energy in the polar cap.

It appears unlikely that the time average of the terms on the LHS of (1.1) and (1.2) would disappear entirely during the six month period of the study. Estimate of these two terms were made using synoptic maps at the beginning and end of the period. The values of the last terms of equations (1.1) and (1.2) could not be evaluated for our study.

There have been some previous studies of the energy balance of the stratosphere, but all have been sketchy and incomplete. White (1954) showed that there was countergradient northward flow of heat in the lower stratosphere. We have verified this result at the 100 mb level using our more complete data. White and Nolan (1959) investigated the term (1.3) in the stratosphere over North America for a five day period. The results suggested the possibility that, on the average, the observed kinetic energy of stratospheric motions may not be replenished by this term (1.3). This transformation term was investigated over the whole northern hemisphere for the six month period. We also evaluated the other terms of (1.1) to determine from whence came the major part of the kinetic energy of the stratosphere.

Since the start of the collection and processing of the data

in the spring of 1959, there have been three other studies at M.I.T. touching on the energy balance of the stratosphere. These three studies drew on two months of data processed for Major C. Jensen by the National Weather Records Center in Ashville, N.C. Approximately 80 of these stations reported at the 100 mb and 50 mb levels during the two months January 1958 and April 1958. In our study we have over 200 stations reporting at these levels giving us much better coverage and detail over the whole hemisphere. As has been pointed out by Hering (1959), the stratospheric flow seems to deviate from its norm for much longer time periods than the tropospheric flow. This means that averages over very long intervals of time have to be taken to approach the true time mean of the various quantities investigated. It is hoped that the averages derived from this study will give something close to the true picture rather than a snapshot over a possibly unrepresentative time period.

Jensen (1960) measured the term (1.3) for the layer 100 mb to 50 mb in January and April 1958, but the results contradict those obtained by White and Nolan mentioned above. With the aid of the better temporal and special coverage we have evaluated this integral.

Working at the 150 mb level and using Jensen's data, Hansrote and Lambert (1960) evaluated the terms $\iint \frac{E\omega}{g} dx\,dy$ and $\iint \frac{\psi\omega}{g} dx\,dy$ between 20°N and 80° N for the month of April 1958. In their study it was necessary to use a layer from 200 to 100 mb to compute the term ω. Since the tropopause is found in this layer, their results cannot necessarily be considered as representative of the stratosphere, especially with the jet stream just below the tropopause.

The severe clear air turbulence associated with the jet stream would lead one to believe that the term $-\iiint \frac{\mathbb{V}\cdot\mathbb{F}}{g}\,dp\,dx\,dy$ would be large in the vicinity of the jet stream, since this term represents the depletion of kinetic energy due to internal "friction" forces.

Roberts (1960) investigated the term $\iint \frac{\varphi\omega}{g}\,dx\,dy$ using Jensen's data for the month of January 1958. For the 250 and 150 mb layers the eddy terms extract kinetic energy from the mass above, while at the 75 mb layer the total kinetic energy of the higher region is increased by the standing and transient eddies of this term. The mean term of this integral was not computed because of difficulties of measuring the mean ω over the region.

Since we use the x, y, p coordinate system throughout this work, we define the word "vertical" to mean along the pressure gradient (Eliassen(1949)). Thus the vertical advection of a quantity means the advection of said quantity across a surface of constant pressure. Unless explicitly stated otherwise, the word "vertical" shall have the above meaning throughout this work. In the same manner "horizontal" shall mean "along a pressure surface".

II Derivation of the Energy Equations for the Stratosphere.

In this section we shall derive the necessary integral equations which will form the bases of our study of the energy balance of the stratosphere.

In the x , y , p coordinate system, x is taken as linear distance with the positive direction towards the east, y is taken as linear distance with the positive direction towards the north and p , the pressure, is taken with the positive direction downwards.

During the IGY the largest amounts of upper air data readily available at uniform reference levels were reported at constant pressure levels. For this reason the equations used below were derived in the x , y , p coordinate system so that the data could be used directly without being converted to the x , y , z coordinate system, where z is upward linear distance.

The horizontal equations of motion in the x , y , p system may be written in the following vector form:

$$\frac{d\mathbb{V}_p}{dt} + f\,\mathbb{K} \times \mathbb{V}_p + \nabla_p \Psi + \mathbb{F} = 0 \qquad (2.1)$$

where $\mathbb{V}_p$ = horizontal component of the velocity of an air parcel on the pressure surface.

$\mathbb{K}$ = unit vertical vector.

$\nabla_p(\;)$ = del operator along the pressure surface.

Ψ = geopotential at the pressure surface.

$\mathbb{F}$ = deceleration due to frictional forces and other forces not included in the other terms.

The first term, $\frac{d\mathbb{V}_p}{dt}$, is the acceleration of the horizontal component of the velocity of an air parcel. The second term $f\,\mathbb{K}\times\mathbb{V}_p$, is the acceleration due to the Coriolis force. The third term, $\nabla_p\,\psi$, is the acceleration due to gravity. $\mathbb{F}$ may be considered as a frictional force. A discussion of this term is contained in Chapter XVI.

Now if we take the dot product of $\mathbb{V}_p$ with equation (2.1) we obtain.

$$\frac{d}{dt}\left(\frac{\mathbb{V}_p\cdot\mathbb{V}_p}{2}\right)+f\mathbb{V}_p\cdot\mathbb{K}\times\mathbb{V}_p+\mathbb{V}_p\cdot\nabla_p\psi+\mathbb{V}_p\cdot\mathbb{F}=0 \quad (2.2)$$

If we define $E\equiv\left(\frac{\mathbb{V}_p\cdot\mathbb{V}_p}{2}\right)$, use the relation $\mathbb{V}_p\cdot\mathbb{K}\times\mathbb{V}_p=0$ and expand the total derivative, (2.2) may be written

$$\frac{\partial E}{\partial t}+\mathbb{V}\cdot\nabla E+\mathbb{V}_p\cdot\nabla_p\psi+\mathbb{V}_p\cdot\mathbb{F}=0 \quad (2.3)$$

where $\nabla(\ )$ is the three-dimensional del operator. Adding $\omega\frac{\partial\psi}{\partial p}$ to $\mathbb{V}_p\cdot\nabla_p\psi$ and subtracting from the rest of the LHS of equation (2.3),

$$\frac{\partial E}{\partial t}+\mathbb{V}\cdot\nabla(E+\psi)-\omega\frac{\partial\psi}{\partial p}+\mathbb{V}_p\cdot\mathbb{F}=0 \quad (2.4)$$

In the x, y, p coordinate system, the continuity equation may be written as $\nabla\cdot\mathbb{V}=0$ or $\frac{\partial\omega}{\partial p}+\nabla_p\cdot\mathbb{V}_p=0$

Now

$$\nabla\cdot(E+\psi)\mathbb{V}=\mathbb{V}\cdot\nabla(E+\psi)+(E+\psi)\nabla\cdot\mathbb{V}=\mathbb{V}\cdot\nabla(E+\psi)$$

which means that (2.4) may be expressed as

$$\frac{\partial E}{\partial t}+\nabla\cdot(E+\psi)\mathbb{V}-\omega\frac{\partial\psi}{\partial t}+\mathbb{V}_p\cdot\mathbb{F}=0 \quad (2.5)$$

Let us now consider the atmosphere north of the latitude circle ϕ_0 and between the two pressure levels p and p_0 $(p < p_0)$. The rate of change of kinetic energy in this polarcap is

$$\iiint \frac{\partial E}{\partial t} \frac{dp}{g} \, dx \, dy = \int_{p}^{p_0} \int_{\phi_0}^{\pi/2} \int_{0}^{2\pi} \frac{\partial E}{\partial t} \frac{a^2 \cos\phi}{g} \, d\lambda \, d\phi \, dp$$

where λ is the longtitude and ϕ the latitude, both measured in **radians**, and a is the radius of the earth, so that $dx = a \cos\phi \, d\lambda$ and $dy = a \, d\phi$.

Taking the triple integral of (2.5) after multiplying through by $1/g$ where $g =$ the acceleration of gravity, we find that the rate of change of kinetic energy is given by

$$\iiint \frac{\partial E}{\partial t} \frac{dp}{g} \, dx \, dy = -\iiint \nabla \cdot (E + \psi) \mathbb{V} \frac{dp}{g} \, dx \, dy + \iiint \omega \frac{\partial \psi}{\partial p} \frac{dp}{g} \, dx \, dy - \iiint \mathbb{V}_p \cdot \mathbb{F} \frac{dp}{g} \, dx \, dy \quad (2.6)$$

In order to simplify the equation, we make the assumption that g is a constant. Application of the divergence theorem to the first integral on the RHS of (2.6) gives:

$$-\iiint \nabla \cdot (E + \psi) \mathbb{V} \frac{dp}{g} \, dx \, dy = \iint E V \frac{dp}{g} \, dx - \iint \frac{(E\omega)_L}{g} \, dx \, dy + \iint \frac{(E\omega)_u}{g} \, dx \, dy + \iint \psi V \frac{dp}{g} \, dx - \iint \frac{(\psi\omega)_L}{g} \, dx \, dy + \iint \frac{(\psi\omega)_u}{g} \, dx \, dy \quad (2.7)$$

where subscript L designates the lower surface, p_0, and u designates the upper surface, p. The first integral on the RHS of (2.7) is the transport of kinetic energy northward across the vertical boundary at latitude ϕ_0.

The second and third terms are the transports of kinetic energy across the lower and upper boundary surfaces. The last three integrals of (2.7) represent the total work done on the polar cap at the boundaries.

Of the two remaining terms on the RHS of (2.6), the last represents the dissipation of kinetic energy due to frictional and other forces.

Thus far we have used only the horizontal equations of motion and the hydrostatic equation for the atmosphere. If we again use the hydrostatic assumption in the form $\frac{\partial \psi}{\partial p} = -\alpha$ and if we consider that the atmosphere satisfies the equation of state for a perfect gas, the remaining term of (2.6) may be written in the alternate forms

$$\iiint \omega \frac{\partial \psi}{\partial p} \frac{dp}{g} dx\, dy = -\iiint \omega \alpha \frac{dp}{g} dx\, dy = -\iiint \frac{R}{g} \frac{\omega T}{p} dp\, dx\, dy$$
$$= -\frac{R}{g} \iiint \omega T\, dx\, dy\, d(\ln p)$$

This same term appears, but with opposite sign, in the equation for the rate of change of potential plus internal energy:

$$\frac{\partial}{\partial t} \iiint (P + I) \frac{dp}{g} dx\, dy = \frac{p_L}{g} \iint \left(\frac{\partial \psi}{\partial t}\right)_L dx\, dy - \frac{p_u}{g} \iint \left(\frac{\partial \psi}{\partial t}\right)_u dx\, dy$$
$$- \iiint \nabla \cdot (C_p T \mathbb{V}) \frac{dp}{g} dx\, dy + \iiint \omega \alpha \frac{dp}{g} dx\, dy + \iiint \frac{dQ}{dt} \frac{dp}{g} dx\, dy \quad (2.8)$$

Equation (2.6) can now be rewritten as

$$\iiint \frac{\partial E}{\partial t} \frac{dp}{g} dx\, dy = \iint E V \frac{dp}{g} dx - \iint \frac{(E\omega)_L}{g} dx\, dy + \iint \frac{(E\omega)_u}{g} dx\, dy$$
$$+ \iint \psi V \frac{dp}{g} dx - \iint \frac{(\psi\omega)_L}{g} dx\, dy + \iint \frac{(\psi\omega)_u}{g} dx\, dy$$
$$- \iiint \omega \alpha \frac{dp}{g} dx\, dy - \iiint \mathbb{V}_p \cdot \mathbb{F} \frac{dp}{g} dx\, dy. \quad (2.9)$$

The method of obtaining numerical values for the integrals will be described in the following chapters.

Equation (2.8) will now be derived. Define the internal and potential energy by:

$\mathcal{J}$ = internal energy per unit mass.

P = potential energy per unit mass.

Let ψ_0 = geopotential at the bottom of the layer considered.

Then $\psi' \equiv \psi - \psi_0$.

Let us consider a small column in the atmosphere with unit cross-section in the horizontal and with top and bottom at the pressure surfaces p_u and p_L . Then the potential energy is given by

$$\int_{p_u}^{p_L} P \frac{dp}{g} = \int_{p_u}^{p_L} \psi \frac{dp}{g} = \int_{p_u}^{p_L} \psi_0 \frac{dp}{g} + \int_{p_u}^{p_L} \psi' \frac{dp}{g} = \frac{\psi_0}{g}(p_L - p_u) + \int_{p_u}^{p_L} \psi' \frac{dp}{g}$$

$$= \frac{\psi_0}{g}(p_L - p_u) + \frac{p_L \psi_L' - p_u \psi_u'}{g} - \int_{\psi_u'}^{\psi_L'} p \frac{d\psi'}{g} \qquad (2.10)$$

Now $\psi_L' = 0$, so using the equation of state and the hydrostatic equation, $\rho\, \partial\psi' = -\partial p$, we obtain

$$\int_{p_u}^{p_L} P \frac{dp}{g} = \frac{\psi_0}{g}(p_L - p_u) - \frac{p_u \psi_u'}{g} + \int_{p_u}^{p_L} RT \frac{dp}{g}$$

Recombination of the first two terms on the RHS gives

$$\int_{p_u}^{p_L} P \frac{dp}{g} = \frac{\psi_L p_L - \psi_u p_u}{g} + \int_{p_u}^{p_L} RT \frac{dp}{g}$$

The time rate of change of the potential energy in a polar cap is then given by

$$\frac{\partial}{\partial t} \iiint P \frac{dp}{g}\, dx\, dy = \frac{p_L}{g} \iint \left(\frac{\partial \psi}{\partial t}\right)_L dx\, dy - \frac{p_u}{g} \iint \left(\frac{\partial \psi}{\partial t}\right)_u dx\, dy$$
$$+ \iiint R \frac{\partial T}{\partial t} \frac{dp}{g}\, dx\, dy \qquad (2.11)$$

The time rate of change of the internal energy of the same polar cap is given by

$$\frac{\partial}{\partial t}\iiint \mathcal{I}\,\frac{dp}{g}\,dx\,dy = \iiint c_v\,\frac{\partial T}{\partial t}\,\frac{dp}{g}\,dx\,dy \tag{2.12}$$

Adding equations (2.11) and (2.12) and using the relation $C_p = R + C_v$

$$\frac{\partial}{\partial t}\iiint (\Phi+\mathcal{I})\frac{dp}{g}\,dx\,dy = \frac{p_L}{g}\iint\left(\frac{\partial\psi}{\partial t}\right)_L dx\,dy - \frac{p_u}{g}\iint\left(\frac{\partial\psi}{\partial t}\right)_u dx\,dy + \iiint c_p\frac{\partial T}{\partial t}\frac{dp}{g}\,dx\,dy$$

Using the first law of thermodynamics for a unit mass of dry air, this last term may be rewritten.

$$\frac{dQ}{dt} = c_p\,\frac{dT}{dt} - \omega\alpha$$

Expanding $\frac{dT}{dt}$ and rearranging

$$C_p\frac{\partial T}{\partial t} = -\nabla\cdot C_pT\,\mathbb{V} + \omega\alpha + \frac{dQ}{dt}$$

Integrating over the polar cap and using the divergence theorem,

$$\iiint C_p\frac{\partial T}{\partial t}\frac{dp}{g}\,dx\,dy = \iint \frac{C_pTV}{g}\,dx\,dp - \iint\left(\frac{C_pT\omega}{g}\right)_L dx\,dy + \iint\left(\frac{C_pT\omega}{g}\right)_u dx\,dy + \iiint \omega\alpha\frac{dp}{g}\,dx\,dy + \iiint\frac{dQ}{dt}\frac{dp}{g}\,dx\,dy \tag{2.13}$$

Hence we get for the final equation of the time rate of change of the potential plus internal energy

$$\frac{\partial}{\partial t}\iiint(\Phi+\mathcal{I})\frac{dp}{g}\,dx\,dy = \frac{p_L}{g}\iint\left(\frac{\partial\psi}{\partial t}\right)_L dx\,dy - \frac{p_u}{g}\iint\left(\frac{\partial\psi}{\partial t}\right)_u dx\,dy + \iint\frac{C_pTV}{g}\,dp\,dx - \iint\left(\frac{C_pT\omega}{g}\right)_L dx\,dy + \iint\left(\frac{C_pT\omega}{g}\right)_u dx\,dy + \iiint\omega\alpha\frac{dp}{g}\,dx\,dy + \iiint\frac{dQ}{dt}\frac{dp}{g}\,dx\,dy \tag{2.14}$$

As with equation (2.9), numerical values of the integrals will be obtained as described in the following chapters.

III Numerical Method of Approximating the Integrals.

We shall apply the technique of Starr & Whito (1952b) for obtaining values of the integrals. Using the following notation where S is a dummy quantity, t is time, p is pressure, λ is longitude and ϕ is latitude:

Time average, denoted by superior bar=

$$\bar{S} \equiv \frac{1}{t-t_0}\int_{t_0}^{t} S\,dt \tag{3.1}$$

Zonal average, denoted by brackets=

$$[S] \equiv \frac{1}{2\pi}\int_{0}^{2\pi} S\,d\lambda \tag{3.2}$$

Meridional average, denoted by braces=

$$\{S\} \equiv \int_{0}^{\pi} S\cos\phi\,d\phi \tag{3.3}$$

Vertical average, denoted by parenthesis=

$$(S) \equiv \frac{1}{p-p_0}\int_{p_0}^{p} S\,dp \tag{3.4}$$

Deviation from the time average, denoted by a prime=

$$S' \equiv S - \bar{S} \tag{3.5}$$

Deviation from the zonal average, denoted by an asterisk=

$$S^* \equiv S - [S] \tag{3.6}$$

Deviation from the meridional average, denoted by double prime=

$$S'' \equiv S - \{S\} \tag{3.7}$$

Deviation from the vertical average, denoted by triple prime=

$$S''' \equiv S - (S) \tag{3.8}$$

As Starr & White (1952b, 1954) and others have shown, for two quantities χ and y we have the following expansion

$$\left(\{[\overline{xy}]\}\right) = \left(\{[\bar{x}]\}\right)\left(\{[\bar{y}]\}\right) + \left(\{[\bar{x}]\}'''\{[\bar{y}]\}'''\right) + \left(\{[\bar{x}]''[\bar{y}]''\}\right) + \left(\{[\bar{x}^*\bar{y}^*]\}\right) + \left(\{[\overline{x'y'}]\}\right) \tag{3.9}$$

But we also see that

$$\left(\left\{\left[\overline{xy}\right]\right\}\right) = \frac{1}{p-p_0}\int_{p_0}^{p}\int_{0}^{\pi}\frac{1}{2\pi}\int_{0}^{2\pi}\frac{1}{t-t_0}\int_{t_0}^{t} xy\, dt\, d\lambda \cos\phi\, d\phi\, dp \tag{3.10}$$

$$\int_{p_0}^{p}\int_{0}^{\pi}\int_{0}^{2\pi}\int_{t_0}^{t} xy\cos\phi\, dt\, d\lambda\, dp\, d\phi = (p-p_0)\, 2\pi\,(t-t_0)\left(\left\{\left[\overline{xy}\right]\right\}\right) \tag{3.11}$$

Thus integrals like (3.11) can be broken down into a number of parts as shown by (3.9). The order of integration in (3.10) is immaterial, but is the essence of the expansion of (3.9).

Each term of equation (3.9) will be explained in detail.

$\left(\left\{\left[\overline{xy}\right]\right\}\right)$ = the mean time average of the total mean interaction of the two quantities through the volume and time.

$\left(\left\{\left[\overline{x}\right]\right\}\right)\left(\left\{\left[\overline{y}\right]\right\}\right)$ = contribution due to the product of the volume means of the two quantities.

$\left(\left\{\left[\overline{x}\right]\right\}'''\left\{\left[\overline{y}\right]\right\}'''\right)$ = contribution of the standing eddies in the vertical. This quantity is the covariance of the $\left\{\left[\overline{x}\right]\right\}$ and $\left\{\left[\overline{y}\right]\right\}$ in the vertical.

$\left(\left\{\left[\overline{x}\right]''\left[\overline{y}\right]''\right\}\right)$ = contribution of the standing eddies due to meridional circulations. This term is the pressure average at each level of the covariance of the quantities $\left[\overline{x}\right]$ and $\left[\overline{y}\right]$.

$\left(\left\{\left[\overline{x}^{*}\,\overline{y}^{*}\right]\right\}\right)$ = contribution of the standing zonal eddies. This quantity is the pressure and meridional average of the covariance of $\overline{x}$ and $\overline{y}$.

$\left(\left\{\left[\overline{x'y'}\right]\right\}\right)$ = contribution of the transient eddies. This quantity is the pressure, meridional and latitudinal average of the covariance of x and y

The procedure for obtaining the terms of (3.9) has been explained before by Starr and White (1952b, 1954), Jensen (1960) and others.

Time means of the quantities were first taken to give $\overline{\chi}$ and $\overline{q}$ for each station at each level. Time covariance, $\overline{\chi' q'}$, were likewise computed. For each level and for each of these quantities a northern hemisphere map was plotted. After analyzing, grid point values were taken at each 10° of longitude and each 5° of latitude from 80° N to the equator. These values were then used to obtain latitudinal averages ie.,$[\overline{\chi}]$, $[\overline{q}]$ and $[\overline{\chi' q'}]$. Latitudinal covariances also gave us $[\overline{\chi}^{*}\overline{q}^{*}]$.

Meridional averages (3,3) were taken of the above quantities giving $\{[\overline{\chi}]\}$, $\{[\overline{q}]\}$, $\{[\overline{\chi' q'}]\}$ and $\{[\overline{\chi}^{*}\overline{q}^{*}]\}$. Meridional spacial covariances also gave us $\{[\overline{\chi}]''[\overline{q}]''\}$.

Vertical averages (3.4) then gave us all of the terms on the RHS of equation (3.9) except the second. The "pressure eddies" of "vertical eddies" were not calculated since there were at most only four pairs of values available to compute the vertical covariance and therefore no confidence could be placed in the statistics.

This gives an outline of the numerical procedure used to evaluate the integrals. For additional details the reader is referred to papers mentioned in this chapter.

IV Method for Computing ω .

As seen in a previous section the quantity $\omega \equiv \frac{dp}{dt}$ plays a major role in the energy equation. Since ω is not one of the values measured, it is necessary to compute it from the quantities which are observed.

Starting with the first law of thermodynamics we obtain, in the pressure coordinate system,

$$\frac{\partial T}{\partial t} + \mathbb{V}_p \cdot \nabla_p T + \omega \frac{\partial T}{\partial t} - \frac{\omega \alpha}{c_p} = \frac{1}{c_p} \frac{dQ}{dt} \tag{4.1}$$

Solving for ω and using the equation of state to express α in terms of p and T ,

$$\omega = \frac{\frac{\partial T}{\partial t} + \mathbb{V}_p \cdot \nabla_p T}{\frac{RT}{c_p p} - \frac{\partial T}{\partial p}} - \frac{\frac{1}{c_p} \frac{dQ}{dt}}{\frac{RT}{c_p p} - \frac{\partial T}{\partial p}} \tag{4.2}$$

Since $\frac{dQ}{dt}$ is not measured on a daily basis, we are not able to evaluate the second term on the RHS of equation (4.2). Therefore we shall evaluate only the first term on the RHS, the adiabatic vertical motions which we denote by ω_c . Thus

$$\omega_c = \frac{\frac{\partial T}{\partial t} + \mathbb{V}_p \cdot \nabla_p T}{\frac{RT}{c_p p} - \frac{\partial T}{\partial p}} \tag{4.3}$$

ω_c is evaluated as a mean quantity between two pressure layers over a given time interval by approximating the differentials by finite differences.

Let $\Delta t = t_f - t_o$ be the time interval and $\Delta p = p_1 - p_2$, the pressure interval. Subscript "1" denotes the lower (higher pressure) layer and subscript "2" denotes the upper layer. A primed quantity denotes the

quantity at the end of the time period, while an unprimed quantity denotes the quantity at the beginning of the time period. Each of the terms on the RHS of (4.3) will be expressed in this notation.

$\frac{\partial T}{\partial t}$ is the mean rate of change of the mean temperature of the pressure layer Δp.

$$\frac{\partial T}{\partial t} = \left(\frac{T_1' + T_2'}{2} - \frac{T_1 + T_2}{2}\right) \Big/ \Delta t = \frac{T_1' + T_2' - T_1 - T_2}{2\Delta t}.$$

$\frac{\partial T}{\partial p}$ is the change of the time mean temperature with respect to pressure.

$$\frac{\partial T}{\partial p} = \frac{1}{2}\left(\frac{T_1 - T_2}{p_1 - p_2} + \frac{T_1' - T_2'}{p_1 - p_2}\right) = \frac{T_1 + T_1' - T_2 - T_2'}{2\,\Delta p}.$$

$\frac{RT}{C_p p}$ is the lapse of temperature with respect to pressure under adiabatic conditions. Here p is the mean pressure for the layer Δp and T is the mean of the temperature over the layer Δp and the time Δt.

$$\frac{RT}{C_p p} = \frac{R}{C_p}\left(\frac{T_1 + T_2 + T_1' + T_2'}{4}\right) \Big/ \left(\frac{p_1 + p_2}{2}\right) = \frac{R}{C_p} \cdot \frac{T_1 + T_2 + T_1' + T_2'}{2(p_1 + p_2)}.$$

Thus we see that these three terms may be approximated by the observed data.

Since individual daily maps were not drawn it was necessary to find other means of measuring the term $\mathbb{V}_p \cdot \nabla_p T$. Using the data at a given station, the geostrophic thermal wind between two pressure levels was used.

If $\mathbb{V}_M \equiv \left(\frac{\mathbb{V}_1 + \mathbb{V}_2}{2}\right)$, $T_M \equiv \left(\frac{T_1 + T_2}{2}\right)$ and the thermal wind

$\mathbb{V}_T = \mathbb{V}_2 - \mathbb{V}_1$, then $\mathbb{V}_M \cdot \nabla_p T_M = \frac{f(p_1 + p_2)}{R\,2(p_1 - p_2)} \mathbb{V}_M \cdot \mathbb{V}_T \times \mathbb{K}$. (4.4)

Let $\mathbb{V}_1 \equiv u_1 \hat{\imath} + v_1 \hat{\jmath}$ and $\mathbb{V}_2 \equiv u_2 \hat{\imath} + v_2 \hat{\jmath}$ then

$$\mathbb{V}_T = (u_2 - u_1)\hat{\imath} + (v_2 - v_1)\hat{\jmath} \quad \text{and} \quad \mathbb{V}_M = \frac{(u_2 + u_1)}{2}\hat{\imath} + \frac{(v_2 + v_1)}{2}\hat{\jmath}$$

$$\begin{aligned}
\mathbb{V}_M \times \mathbb{V}_T &= \left(\frac{u_2 + u_1}{2}\right)(v_2 - v_1)\,\mathbb{K} - \left(\frac{v_2 + v_1}{2}\right)(u_2 - u_1)\,\mathbb{K} \\
&= \frac{\mathbb{K}}{2}\left(u_2 v_2 - u_2 v_1 + u_1 v_2 - u_1 v_1 - \{ v_2 u_2 - v_2 u_1 + v_1 u_2 - v_1 u_1 \}\right) \\
&= \frac{\mathbb{K}}{2}\left(2 u_1 v_2 - 2 u_2 v_1\right) = \mathbb{K}(u_1 v_2 - u_2 v_1)
\end{aligned}$$

Since $\mathbb{V}_M \cdot \mathbb{V}_T \times \mathbb{K} = \mathbb{K} \cdot \mathbb{V}_M \times \mathbb{V}_T = u_1 v_2 - u_2 v_1$ we obtain

$$\mathbb{V}_p \cdot \nabla_p T = \mathbb{V}_M \cdot \nabla_p T_M = \frac{f(p_1 + p_2)}{2R(p_1 - p_2)}(u_1 v_2 - u_2 v_1)$$

if we approximate $\mathbb{V}_p \cdot \nabla_p T$ by $\mathbb{V}_M \cdot \nabla_p T_M$ through the layer Δp . Since ω_c is computed over a time interval we will take the average $\mathbb{V}_p \cdot \nabla_p T = \frac{f}{4R}\left(\frac{p_1 + p_2}{p_1 - p_2}\right)(u_1 v_2 - u_2 v_1 + u_1' v_2' - u_2' v_1')$

Combining the finite difference forms we obtain the difference equation for ω_c :

$$\omega_c = \frac{\dfrac{T_1' + T_2' - T_1 - T_2}{\Delta t} + \dfrac{\Omega \sin\phi\,(p_1 + p_2)}{R(p_1 - p_2)}(u_1 v_2 - u_2 v_1 + u_1' v_2' - u_2' v_1')}{R/c_p \cdot (T_1 + T_2 + T_1' + T_2')/(p_1 + p_2) - (T_1 + T_1' - T_2 - T_2')/(p_1 - p_2)} \tag{4.5}$$

where Ω = angular velocity of the earth's rotation and ϕ = latitude of the station.

Because of the diurnal errors in the temperature measuring elements of the radiosondes, a twenty-four hour time period was used for calculating ω_c The wisdom of this choice was later born out when Major Jensen's (1960) vertical motions were calculated. His highest Δp layer and our lowest Δp layer were the same, and we had a number of the same stations for the same months. The twelve hour values of ω_c usually

alternated in sign, but the average of two successive twelve hour values was almost exactly equal to our twenty-four hour values of ω_ℓ . The discrepancies were due to round off errors and different methods of computing the term $\mathbb{V} \cdot \nabla T$.

A personal communication from Dr. T. Murakami indicated that a twenty-four hour time period is a better time period to use than twelve hours.

As mentioned before the diabatic part of equation (4.2) cannot be measured on a daily basis. This term is discussed in Chapters VIII and XVI.

A primed quantity in this chapter has denoted the quantity at the end of the time period. This definition is used only in this chapter and should not be confused with the definition given in Chapter III and used in all other chapters.

V Data Used in the Study.

Since the atmosphere is its own best model we have used the data from actual soundings to evaluate the integrals. The equations were derived in the x, y, p coordinate system because the observations were taken with pressure as one of the basic measurements. The heights of the pressure surfaces were derived from the temperature and pressure observations, and hence also have inaccuracies due to computational, observational and instrumental errors.

Our object was to obtain as many stations as possible over the northern hemisphere, which reported daily values of the height of the pressure surface, the temperature and the wind at 100, 50, 30 and 10 mb. These four levels were standard reporting levels in the stratosphere for the WMO stations during the IGY. Only actual data were used. There were no extrapolations and no substitutions of alternate stations.

This work was started in January 1959 with the expectation of early receipt of the IGY microcards. Repeated delays in the production of the microcards forced us to begin by using other data sources. The decision was made to use those data sources which were available for a study of the first six months of the IGY and to await the receipt of the microcards before beginning the study of the year 1958.

A hemispheric network of stations was chosen so as to give good and fairly complete coverage. Only stations reporting winds, temperatures and heights were considered. Of these, those which reported two or more times a day were given preference. Because of the large number of these preferred stations in certain areas, some were eliminated on the grounds of lower frequency of observations in the stratosphere.

The U.S. Weather Bureau Daily Upper Air Bulletin (DUAB) supplied the needed data over North America and for the U.S. Air Force station bases throughout the world. For the IGY, 100 50, 30 and 10mb surfaces were the WMO mandatory levels in the stratosphere, but the DUAB used 25 mb in place of 30 mb. Additional data from U.S. military bases throughout the world were obtained on microfilm from the National Weather Record Center in Ashville, North Carolina. The microfilm provided data at the WMO mandatory levels. Microcards containing a majority of the remaining selected stations arrived before station data from the above sources were completely compiled and processed.

Data from the People's Republic of China were not available through the WMO since the People's Republic of China did not participate in the International Geophysical Year. However, data had been obtained by radiointercept and were made available through the National Weather Record Center, Ashville, North Carolina. The Chinese data above 100 mb were all but useless for our purposes.

In areas of poor coverage it was necessary to use stations with only one run a day and stations with rather meager stratospheric data.

A few stations in North Africa and India used 100, 60 and 20 mb as mandatory levels and took observations at 06Z and 18Z instead of 00Z and 12Z, the standard observation times.

No stations on the South American continent north of the equator were available before the end of the computation stage. Even though there were a surprising number of rawinsonde stations in North Africa, only seven reported stratospheric data frequently enough to be used. It was

also difficult to find good stations in Asia south of 25°North. The Russian stations were good up to 30 mb, 10 mb reports being almost non-existent. The lack of good stratospheric reporting stations in northernmost Russia made the analysis of the polar regions difficult. The complete absence of reports from Mexico and to the west in the Pacific made this region as difficult to analyze as the Central Atlantic Ocean.

Two hundred and eleven stations listed in Table 5.1 by International Index number and shown in Plate 1 were used. Over the People's Republic of China the station numbers are those used by the People's Republic of China. All other station numbers are WMO index numbers.

No extrapolations were made because there is always the possiblity that these extrapolations might bias the statistics. On the other hand we are fully aware that soundings are usually terminated at lower levels when the wind speed is very large or when the air is very cold. Since the stations were selected on a basis of good performance at 100 mb and 50 mb, any bias due to the above two factors should be small. Above 50 mb we probably have a bias due to lighter winds and warmer temperatures at lower levels.

Table 5.1. Stations used in this study.

INDEX NUMBER	NAME	LATITUDE	LONGTITUDE
02-077	Stockholm	59.4N	18.0E
03-005	Lerwick	60.1N	01.2W
03-171	Leuchars	56.4N	02.9W
03-322	Fazakerly	53.6N	02.9W
03-496	Hemsby	52.7N	01.7E
03-774	Crawley	51.1N	00.2W
03-808	Camborne	50.2N	05.3W
04-018	Keflavik	64.0N	22.6W
04-202	Thule	76.5N	68.8W
04-270	Narssarssuaq	61.2N	45.4W
04-310	Nord	81.6N	16.7W
06-180	Copenhagen	55.6N	12.7E
06-610	Payerne	46.8N	07.0E
07-170	Chaumont	48.1N	05.0E
08-509	Lajes	38.8N	27.1W
10-610	Bitburg	50.0N	06.5E
10-739	Stuttgart	48.8N	09.2E
11-035	Wienhoke	48.3N	16.4E
16-080	Milano	45.5N	09.3E
16-239	Rome	41.8N	12.6E
16-320	Brindisi	40.7N	18.0E
17-606	Nicosia	35.2N	33.3E

Table 5.1 . (continued)

INDEX NUMBER	NAME	LATITUDE	LONGTITUDE
22-113	Murmansk	69.0N	33.1E
22-522	Kem-Port	65.0N	34.8E
22-550	Arkhangelsk	64.6N	40.5E
23-472	Turukhansk	65.8N	88.0E
23-955	Aleksandrovskoe	60.4N	77.9E
24-266	Verkhoyansk	67.6N	133.4E
24-641	Viliuysk	63.8N	121.6E
24-688	Oimyakon	63.3N	143.2E
24-817	Erbogatchen	61.3N	108.0E
24-959	Yakutsk	62.1N	129.8E
25-563	Anadyr	64.8N	177.6E
25-703	Seimtchan	62.9N	152.4E
25-954	Korf/Tilichiki	60.4N	166.0E
26-063	Leningrad	60.0N	30.3E
26-702	Kaliningrad	54.7N	20.6E
26-850	Minsk	53.9N	27.5E
27-037	Vologda	59.3N	39.9E
27-196	Kirov	58.6N	49.6E
27-612	Moscow	55.8N	37.6E
28-440	Sverdlovsk	56.8N	60.6E
28-698	Omsk	54.9N	73.4E
28-900	Kwibishev	53.2N	50.5E
28-952	Kustanay	53.2N	63.6E
29-574	Krasnoiarsk	56.0N	92.9E

Table 5.1 (continued)

INDEX NUMBER	NAME	LATITUDE	LONGTITUDE
29-634	Novosibirsk	55.0N	82.9E
30-230	Kirensk	57.8N	108.1E
30-554	Troitskiy Priisk	54.6N	113.1E
30-758	Thita	52.0N	113.1E
30-935	Krasny Tchikoi	50.4N	108.8E
31-004	Aldan	58.6N	125.4E
31-168	Aian	56.5N	138.2E
31-510	Blagoveshchensk	50.3N	127.5E
31-735	Khabarovsk	48.5N	135.2E
31-960	Vladivostok	43.1N	131.9E
32-061	Aleksandrovsk	50.9N	142.2E
32-150	B'Elan	46.9N	142.7E
32-540	Petropalovsk-Kamchatsky	53.0N	158.8E
33-345	Kyev	50.4N	30.4E
33-393	Lvov	49.8N	24.0E
33-837	Odessa	46.5N	30.6E
33-946	Simferopol	45.0N	34.0E
34-300	Kharkov	49.9N	36.3E
34-560	Voroponovo	48.7N	44.4E
35-229	Aktiubinsk	50.3N	57.2E
35-394	Karaganda	49.8N	73.1E
35-700	Gurev	47.1N	51.9E
35-746	Aralskoyemore	46.8N	61.7E
35-796	Balkahash	46.9N	75.0E

Table 5.1 (continued)

INDEX NUMBER	NAME	LATITUDE	LONGITUDE
36-177	Semipalatinsk	50.4N	80.2E
36-870	Alma-Ata	43.2N	76.9E
37-549	Tbilisi	41.7N	45.0E
38-457	Tashkent	41.3N	69.3E
38-880	Ashkhabad	38.0N	58.3E
40-427	Bahrain	26.3N	50.6E
40-597	Aden	12.8N	45.0E
40-648	Habbaniya	33.4N	43.6E
42-182	New Delhi	28.6N	77.2E
43-279	Madras	13.0N	80.2E
43-466	Colombo	6.9N	79.9E
45-004	Hong Kong	22.3N	144.2E
46-697	Taoyuan	25.0N	121.2E
47-132	Mosulpo	33.2N	126.2E
47-412	Sapporo	43.0N	141.3E
47-600	Wajima	37.4N	136.9E
47-646	Tateno	36.0N	140.1E
47-778	Shionomisah	33.4N	135.8E
47-909	Naze	28.4N	129.5E
47-931	Okinawa	26.4N	127.8E
47-963	Torishima	30.5N	140.3E
48-455	Bangkok	13.7N	100.5E
50-838	Wulanhot	46.2N	122.0E

Table 5.1 (continued)

INDEX NUMBER	NAME	LATITUDE	LONGITUDE
51-828	Hotien	37.1N	79.9E
52-418	Tunhwang	40.1N	94.8E
52-818	Karmu	36.2N	94.6E
54-102	Hsilinhot	43.8N	116.1E
55-591	Lhasa	29.7N	91.0E
57-036	Sian	34.2N	108.9E
57-816	Kweiyang	26.6N	106.7E
58-367	Shanghai	31.2N	121.4E
60-119	Port Lyautey	34.3N	06.6W
61-642	Dakar	14.7N	17.4W
62-011	Tripoli	32.9N	13.3E
62-053	Benina	32.1N	20.3E
62-721	Khartoum	15.6N	32.6E
64-910	Douala	4.0N	9.7E
65-578	Abidjan	5.2N	3.9W
70-026	Barrow	71.3N	156.8W
70-086	Barter Island	70.1N	143.7W
70-200	Nome	64.5N	165.4W
70-219	Bethel	60.8N	161.7W
70-231	McGrath	63.0N	155.6W
70-261	Fairbanks	64.8N	147.9W
70-308	St Paul Island	57.2N	170.2W
70-316	Cold Bay	55.2N	162.7W

Table 5.1 (continued)

INDEX NUMBER	NAME	LATITUDE	LONGITUDE
70-350	Kodiak	57.8N	152.5W
70-361	Yakutat	59.5N	139.7W
70-398	Annette Island	55.0N	131.6W
70-409	Attu	52.8N	173.2E
70-454	Adak	51.9N	176.6W
72-201	Key West	24.6N	81.8W
72-208	Charleston, S.C.	32.9N	80.0W
72-211	Tampa	28.0N	82.5W
72-235	Jackson	32.3N	90.2W
72-250	Brownsville	25.9N	97.5W
72-259	Fort Worth	32.8N	97.0W
72-261	Del Rio	29.3N	100.9W
72-265	Midland	32.9N	102.2W
72-270	El Paso	31.8N	106.4W
72-274	Tuscon	32.1N	111.0W
72-290	San Diego	32.7N	117.2W
72-308	Norfolk	36.9N	76.2W
72-327	Nashville	36.1N	86.7W
72-386	Las Vegas	36.1N	115.2W
72-405	Washington, D.C.	38.8N	77.0W
72-456	Topeka	39.1N	95.6W
72-469	Denver	39.8N	104.9W
72-493	Oakland	37.7N	122.2W

Table 5.1 (continued)

INDEX NUMBER	NAME	LATITUDE	LONGITUDE
72-518	Albany	42.8N	73.8W
72-537	Detroit	42.4N	83.0W
72-572	Salt Lake City	40.8N	112.0W
72-597	Medford	42.4N	122.9W
72-645	Green Bay	44.5N	88.1W
72-655	St. Cloud	45.6N	94.2W
72-662	Rapid City	44.2N	103.0W
72-712	Caribou	46.9N	68.0W
72-734	Sault Ste. Marie	46.5N	84.4W
72-747	International Falls	78.6N	93.4W
72-764	Bismark	46.8N	100.8W
72-768	Glasgow	48.2N	106.6W
72-785	Spokane	47.6N	117.5W
72-798	Tatoosh Island	48.4N	124.7W
72-815	Harmon	48.5N	58.6W
72-816	Goose Bay	53.3N	60.4W
72-836	Moosonee	51.3N	80.6W
72-848	Trout Lake	53.8N	89.9W
72-879	Edmonton	53.6N	113.5W
72-896	Prince George	53.9N	122.7W
72-906	Fort Chimo	58.1N	68.4W
72-913	Churchill	58.8N	94.1W
72-917	Eureka	80.0N	85.9W

Table 5.1 (continued)

INDEX NUMBER	NAME	LATITUDE	LONGITUDE
72-924	Resolute Bay	74.7N	95.0W
72-926	Baker Lake	64.3N	96.0W
72-934	Fort Smith	60.0N	112.0W
72-938	Coppermine	67.8N	115.1W
72-945	Fort Nelson	58.8N	122.6W
74-043	Norman Wells	65.3N	126.8W
74-051	Sachs Harbor	72.0N	124.7W
74-074	Isachsen	78.8N	103.5W
74-082	Alert	82.5N	62.3W
76-644	Merida	21.0N	89.5W
78-016	Kindley	32.4N	64.7W
78-063	Gold Rock Creek	26.6N	78.3W
78-118	Turks Island	21.5N	71.1W
78-397	Kingston	17.9N	76.8W
78-501	Swan Island	17.4N	83.9W
78-526	San Juan	18.5N	66.1W
78-663	San Salvador	13.7N	89.1W
78-806	Albrook	9.0N	79.6W
78-866	Juliana Airfield	18.0N	63.1W
78-967	Trinidad	10.7N	61.6W
78-988	Willemstad	12.2N	69.0W
80-001	St. Andrews	12.5N	81.7W
91-066	Midway	28.2N	177.4W
91-115	Iwo Jima	24.8N	141.3E

Table 5.1 (continued)

INDEX NUMBER	NAME	LATITUDE	LONGITUDE
91-131	Marcus Island	24.3N	154.0E
91-165	Lihue	22.0N	159.4W
91-217	Guam	13.6N	144.8E
91-245	Wake Island	19.3N	166.7E
91-250	Eniwetok	11.3N	162.3E
91-275	Johnson Island	16.7N	169.5W
91-285	Hilo	19.7N	155.1W
91-334	Truk	7.5N	151.8E
91-348	Ponape	7.0N	158.2E
91-366	Kwajalein	8.7N	167.7E
91-376	Majuro	7.1N	171.4E
91-408	Koror	7.4N	134.5E
91-413	Yap	9.5N	138.1E
91-489	Christmas Island	2.0N	157.4W
91-700	Canton Island	2.8S	171.7W
98-327	Clark	15.2N	120.6E
	Arctic "A"	83 N	168 W
	NP-6	85 N	179 W
4YB	Ship Atlantic B	56.5N	51.0W
4YC	Ship Atlantic C	52.8N	35.5W
4YD	Ship Atlantic D	44.0N	41.0W
4YE	Ship Atlantic E	35.0N	48.0W
4YN	Ship Pacific N	30 N	140 W
4YP	Ship Pacific P	50 N	145 W
4YV	Ship Pacific V	34 N	164 E

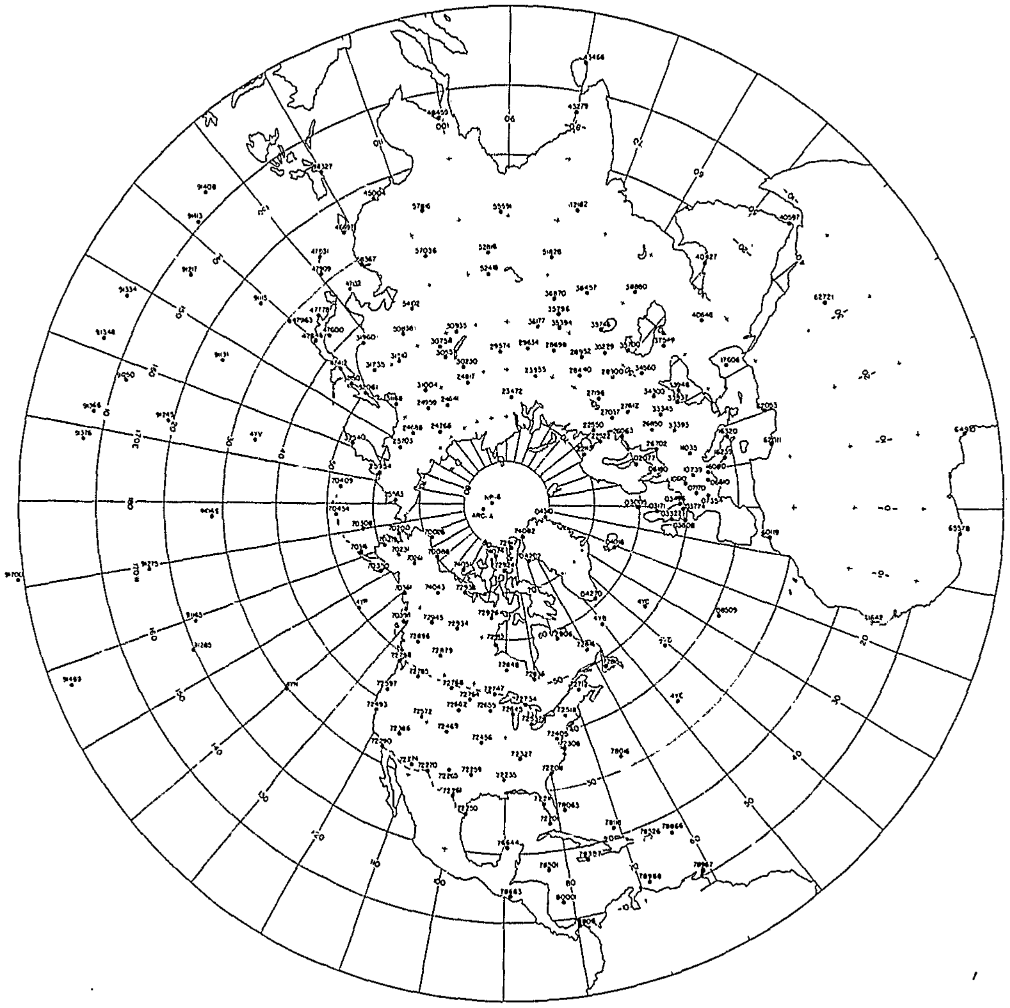


Plate 1. Stations, by international index number, used as source for stratospheric data.

VI Methods of Reducing the Data.

The amount of data handling was large enough to require a high-speed computer. An analysis of the computational problems and of the availability of, and initial processing of, the data showed that for our purposes the LGP-30 High-Speed Computer (Royal-McBee Co.) was the best computer available at the time. The LGP-30 and peripheral equipment were delivered in the spring of 1959.

Punched paper tapes were used as input. Each tape contained one month of data from one station for either the 00Z of 12Z observation time. The proofread tapes were first used to obtain monthly averages of the heights, temperatures, westerly components and southerly components of the winds at the four levels. Standard deviations of the items were also computed. Inspection of these quantities pointed out the gross mistakes in the data. This program proved a most effective filter for detecting erroneous data.

The decision to use three-month periods in this study was determined to some extent by the internal memory of the LGP-30. The use of monthly periods was ruled out because of the large number of maps which would be necessary.

For each station three months of data for one time (either 00Z or 12Z) were run through the Stratospheric Flux Program (SFP). Table 6.1 lists the various quantities computed by the SFP and gives an example of the printout from the computer. As explained in the previous chapter, observations were available at 25mb instead of 30 mb for some stations, hence the statistics of these stations were computed for the 25 mb level.

Observations of total amounts of ozone were available on a daily basis for about thirty stations. Studies of ozone using these data have been reported on by R.E. Newell (1961). For the remaining stations the solar index k_p was used instead of ozone. Some stations reported wind speed in knots rather than meters per second, so the units of U and V are knots $x10^{-1}$ for these stations.

From the data tapes other programs computed station values for the $\overline{\omega_c}$, $\overline{C^2}$, $\overline{\omega_c' T'}$, $\overline{\omega_c' H'}$, $\overline{\omega_c' C^2}$ and $\overline{V' C^2}$ maps.

The time means of the individual quantities and the covariances were obtained from various computer programs and the data tapes. These values were then plotted on northern hemisphere maps. After the maps were analyzed, grid point values were taken at every ten degrees of longitude and every five degrees of latitude from 80°N to the equator. The grid point values were put onto punched paper tape so that latitudinal means could be obtained and spacial correlations between the grid point values of two maps could be computed by the high-speed computer.

The number of reporting stations on some of the higher level maps was too small to permit analysis of the maps. Using the "buckshot"method, values were obtained from these maps. The buckshot method consists of taking the average of all of the values within a given latitude belt. This average is then taken as the mean for the latitude belt. This assumes that the stations are randomly scattered, eliminating any bias. Both grid point method and the buckshot method were used on one of the higher level maps where the analysis was rather free but not difficult. The agreement between the values produced by the two methods was satisfactory showing that

Table 6.1 Sample wind, temperature, and pressure-height statistics at 100, 50, 30 and 10 mb pressure levels.

EXPLANATION OF SYMBOLS

1. L — Pressure level (100, 50, 30, 10 mb)

2. I — Item: *

oo	ozone	statistics
oh	ozone-pressure height	"
ot	" -temperature	"
ou	" -zonal wind	"
ov	" -meridional wind	"
hh	pressure height	"
ht	" " -temperature	"
hu	" " -zonal wind	"
hv	" " -meridional wind	"
tt	temperature	"
tu	" -zonal wind	"
tv	" -meridional wind	"
uv	meridional-zonal wind	"

Units:
- o cm x 10^{-3}
- h meters x 10
- t $^{\circ}$C x (-10^{-1})
- u meters sec^{-1} x 10^{-1}
- v meters sec^{-1} x 10^{-1}

3. N — Number of observations

4. $\bar{X}$ — Mean of first item under item column.
$\bar{X} = \frac{1}{N}\sum_{i=1}^{N} X_i$ units of X

5. $\bar{Y}$ — Mean of second item under item column.
$\bar{Y} = \frac{1}{N}\sum_{i=1}^{N} Y_i$ units of Y

6. SX — Standard deviation of first item under item column.
SX $=\sqrt{\sum_{i=1}^{N}(X_i-\bar{X})^2}$ units of X

7. SY — Standard deviation of second item under item column.
SY $=\sqrt{\sum_{i=1}^{N}(Y_i-\bar{Y})^2}$ units of Y

8. CXY — Covariance of the first and second items under item column.
CXY $= \overline{XY} - \bar{X}\bar{Y}$ units of XY

9. RXY — Correlation coefficient between the first and second item under item column.
RXY $= \frac{CXY}{(SX)(SY)}$ units x 10^{-3}

Note: The decimal point is always immediately to the right of numbers. (example: -0740 = -0740.)

* The solar index k_p was used instead of ozone for those stations which did not take observations of total ozone.

L	I	N	$\bar{X}$	$\bar{Y}$	SX	SY	CXY	RXY
100mb	oo	0092	0021		0012			
	oh	0086	0022	0661	0012	0010	0000009	066
	ot	0085	0022	0709	0012	0030	0000076	191
	ou	0084	0022	0080	0012	0101	0000397	305
	ov	0084	0022	-0036	0012	0073	-0000108	-114
	hh	0086	0661		0010			
	ht	0085	0661	0709	0010	0030	0000062	189
	hu	0084	0661	0080	0010	0101	-0000532	-491
	hv	0084	0661	-0036	0010	0073	0000043	054
	tt	0085	0709		0030			
	tu	0084	0710	0080	0030	0101	-0000374	-119
	tv	0084	0710	-0036	0030	0073	0000231	100
	uv	0084	0080	-0036	0101	0073	-0000961	-128
50mb	oo	0083	0022		0012			
	oh	0083	0022	1082	0012	0010	0000001	007
	ot	0083	0022	0584	0012	0021	-0000049	-177
	ou	0082	0022	-0046	0012	0091	0000320	269
	ov	0082	0022	0010	0012	0035	-0000061	-131
	hh	0083	1082		0010			
	ht	0083	1082	0584	0010	0021	-0000073	-316
	hu	0082	1082	-0046	0010	0091	-0000294	-297
	hv	0082	1082	0010	0010	0035	0000041	106
	tt	0083	0584		0021			
	tu	0082	0584	-0046	0021	0091	-0000958	-477
	tv	0082	0584	0010	0021	0035	-0000151	-192
	uv	0082	-0046	0010	0091	0035	-0000205	-062
25mb	oo	0073	0023		0013			
	oh	0073	0023	1408	0013	0013	0000010	059
	ot	0073	0023	0516	0013	0020	-0000025	-090
	ou	0073	0023	-0094	0013	0053	0000167	233
	ov	0073	0023	0012	0013	0036	-0000013	-027
	hh	0073	1408		0013			
	ht	0073	1408	0516	0013	0020	-0000199	-0724
	hu	0073	1408	-0094	0013	0053	-0000206	-294
	hv	0073	1408	0012	0013	0036	0000069	145
	tt	0073	0516		0020			
	tu	0073	0516	-0094	0020	0053	0000008	007
	tv	0073	0516	0012	0020	0036	-0000072	-095
	uv	0073	-0094	0012	0053	0036	-0000703	-364
10mb	oo	0030	0027		0013			
	oh	0030	0027	2134	0013	0018	-0000018	-070
	ot	0030	0027	0419	0013	0034	0000008	017
	ou	0030	0027	-0090	0013	0068	0000154	162
	ov	0030	0027	0021	0013	0023	-0000090	-279
	hh	0030	2134		0018			
	ht	0030	2134	0419	0018	0034	-0000552	-0860
	hu	0030	2134	-0090	0018	0068	-0000446	-354
	hv	0030	2134	0021	0018	0023	0000087	203
	tt	0030	0419		0034			
	tu	0030	0419	-0090	0034	0068	0000887	373
	tv	0030	0419	0021	0034	0023	-0000103	-127
	uv	0030	-0090	0021	0068	0023	0000226	143

mosulpo seasonal july-sept 57 00z solar 9 20 60 r.a.b.

the buckshot method does have some validity for our work.

The author is grateful to Dr. Takio Murakami for analyzing maps of the quantities $\overline{T}$, $\overline{H}$, $\overline{u}$, $\overline{v}$, $\overline{T'v'}$, $\overline{H'v'}$ and $\overline{u'v'}$

In analyzing the other maps the author did not make an effort to strive for consistency between the 00Z maps and the 12Z maps. The maps were analyzed independently so as to give an indication of the variations due to analysis and due to diurnal effects. An attempt was made to reanalyze two maps to obtain better agreement between 00Z and 12Z latitudinal means. This could not be done without violating large amounts of data on one map or the other.

The Stratosphere Flux Program was written by Prof. E. Lorenz, Dr. Robert M. White and the author. The program for obtaining the monthly statistics, programs for obtaining ω_c and the ω_c statistics, programs for obtaining wind statistics and programs for obtaining the required statistics from the grid point tapes were written by the author. All of the programs were written in fixed point to reduce the time required on the computer. Special input programs were devised by the author to pinpoint errors in the input tapes.

VII Discussion of Maps of Time - Mean Quantities.

The adiabatic part of the vertical motion computed using two layers was assigned to the intermediate level. Thus the 100 mb and 50 mb data were used to obtain a mean adiabatic vertical motion which was assigned to the 75 mb level. The similarity of the patterns of the 00Z and 12Z maps and of the statistics derived from these maps show that the mean values are above the noise level. The dissimilarities could be due to diurnal effects in the adiabatic vertical motion. The reduced number of soundings reaching higher levels was the cause of greater disagreement between the 00Z and 12Z maps.

The adiabatic vertical motions are much larger during the winter period.

In Chapter VIII we show that the major features of the maps of the mean adiabatic vertical motions represent the major feature of the mean true vertical motion, $\frac{dp}{dt}$, so certain comments can be made regarding the actual vertical motion. Most of the activity seems to take place in higher latitudes during the winter period with considerable upward motion over Alaska and North Atlantic and sinking motion over Canada and the White Sea. The division over the Kuril Islands of strong sinking motion to the south and rising motion to the north is probably associated with the jet stream.

A break down of the quantity $\overline{c^2}$, the mean square horizontal wind speed, into its u and v components gives

$$\overline{c^2} = \overline{u}^2 + \overline{v}^2 + \overline{u'^2} + \overline{v'^2} \qquad (7.1)$$

But we have $\overline{u'^2} = S(u)^2$ and $\overline{v'^2} = S(v)^2$, so

$$\overline{C^2} = \overline{U}^2 + \overline{V}^2 + S(u)^2 + S(v)^2 \qquad (7.2)$$

Investigation of the individual terms shows that $\overline{V}^2 << \overline{U}^2$ so $\overline{V}^2$ could be dropped in any computation of $\overline{C^2}$ from equation (7.2). The standard deviations of U and V are comparable and they cannot be dropped in relation to the term $\overline{U}^2$.

Since grid point values of $S(u)$ and $S(V)$ were not available, the $\overline{C^2}$'s were computed directly from the data. When the grid point values of $S(u)$ and $S(V)$ become available it is suggested that $\overline{C^2}$ be computed by the use of equation (7.2) and that these values of $\overline{C^2}$ be compared with those obtained by our method.

The maps of $\overline{C^2}$ were used to obtain the kinetic energy of the horizontal winds. The kinetic energy of the vertical wind is negligible compared to the horizontal part. This can easily be shown by including $\overline{W}^2$ and $\overline{W'^2}$ in equation (7.1) and then considering orders of magnitude of the terms. $\overline{C^2}$ is non-negative since there are no complex values of the wind speed. It will be noted that $\overline{C^2}$ is positive everywhere on the maps since, even if both $\overline{U}^2$ and $\overline{V}^2$ are zero for the period, the standard deviations are not zero.

Direct observations of the wind speed at 75 mb and 40 mb were not readily available. Values at 75 mb were obtained by taking the average of the 100 mb and 50 mb wind speeds. 40 mb wind speeds were taken as the average of the 50 mb and 30 mb speeds. It was recognized that this was only one of the many arbitrary ways of interpolating for the speeds.

The presence of the jets which seems to come from the southern hemisphere into south western Africa and then across North Africa is not

clearly documented. Even though there were only a few reporting stations in North Africa, as seen in Plate 1, the seven stations did indicate the presence of this jet. Tiros pictures have shown evidence of a jet stream across the Red Sea.

Observations taken at the 10 mb level show that the jet seems to increase and move further towards the pole than at 30 mb. The 10 mb data were not available over Russia.

The mean heights around the latitude circle of the 100, 50 and 30 mb surfaces for the first of July, last of September and last of December 1957 are shown in Figure 7.1. These heights were obtained from maps published by the U.S. Weather Bureau. The greatest changes take place in the northern latitudes at all three levels. During the first part of July we find mean westerlies at 100 mb and mean easterlies at 50 and 30 mb. By the end of September the westerlies have become stronger at 100 mb and have made their appearance at northern latitudes at 50 mb whereas at 30 mb the westerlies appear only north of 60° N. By the end of December at 100 mb we find stronger mean westerlies, as at the 50 and 30 mb levels. The slope of the mean height profile with latitude indicates that the westerlies increase with height and become stronger towards the pole with height. This is also confirmed by inspection of the daily Weather Bureau maps. The 3 month mean maps of the heights of the 100 mb, 50 mb and 30 mb surfaces have been published by Dr. Murakami, (1962). These maps indicate the same features as mentioned above, however the Weather Bureau maps make the changes more striking since they depict daily conditions rather than mean conditions

for the three month periods.

The changes in the temperatures at these levels are shown in Figures 7.2, 7.3 and 7.4. At 100 mb we find a definite cooling north of 45°N while there are indications of a warming from approximately 45°N to 20°N as we progress from July through December. At the 50 mb level we find that there is definite cooling north of 40°N whereas there is first a warming from July through September and then a cooling from the end of September through the end of December from 40°N through 15°N.

Again at 30 mb we find cooling from the north pole to 35°N from July through December. Between 35°N and 5°N there is a warming from July through September followed by a cooling from the end of September through December just as found at the 50 mb level.

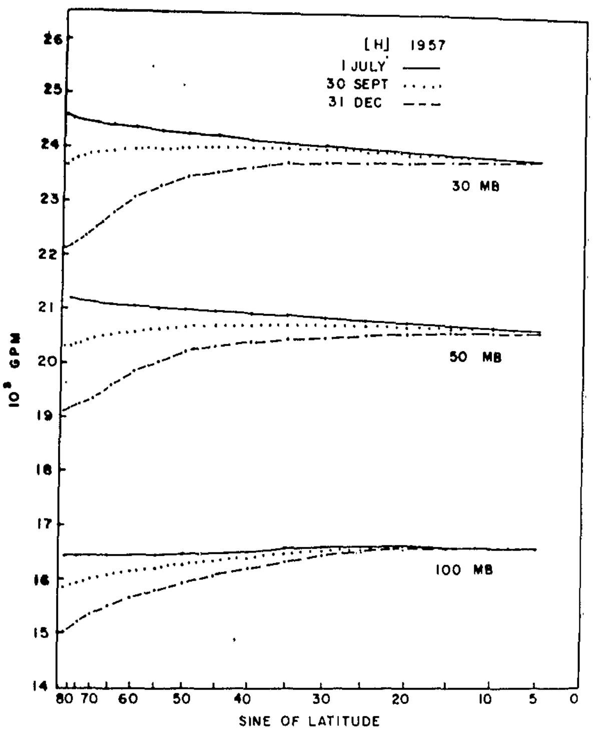


Figure 7.1

Zonal mean heights of the 100,50 and 30 mb surfaces.

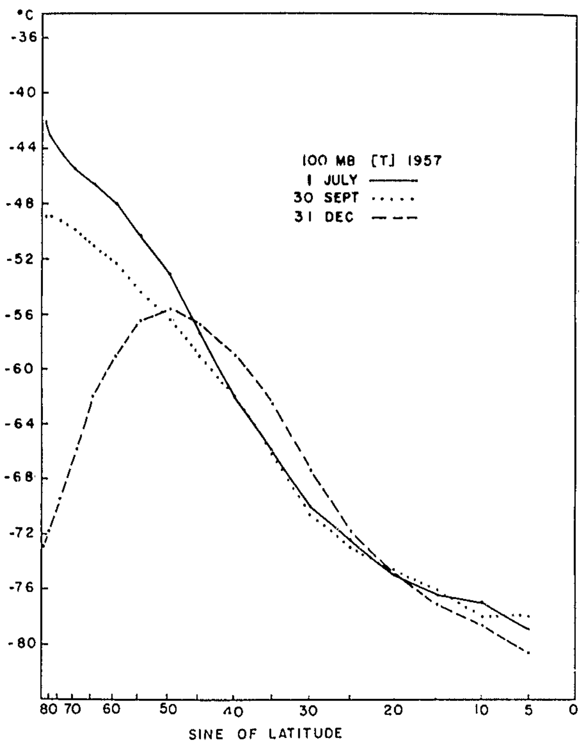


Figure 7.2

Zonal mean temperatures

of the 100 mb surface.

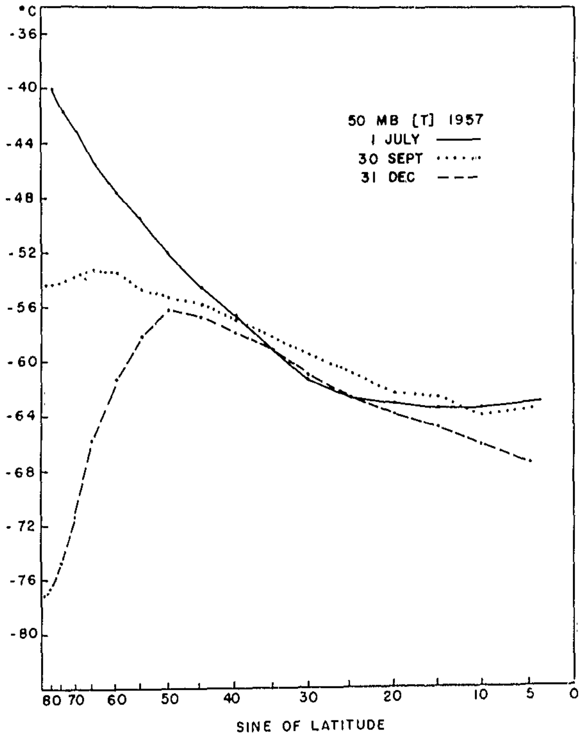


Figure 7.3

Zonal mean temperatures

of the 50 mb surface.

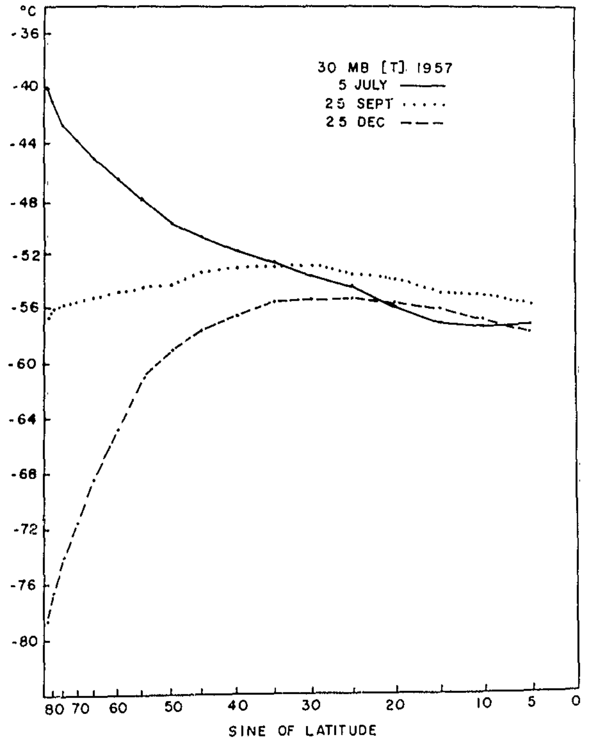


Figure 7.4

Zonal mean temperatures

of the 30 mb surface.

VIII Methods of Obtaining Mean Meridional Velocities

An indirect method of obtaining the mean horizontal meridional motion is by using the mean vertical meridional motion and the continuity equation. The method and an application of the method to our data are discussed.

The total rate of change of temperature of a moving particle can be expressed as

$$\frac{dT}{dt} = \frac{\partial T}{\partial t} + \mathbb{V}_p \cdot \nabla_p T + \omega \frac{\partial T}{\partial p} = \frac{RT}{c_p p}\omega + \frac{1}{c_p}\frac{dQ}{dt} \qquad (8.1)$$

Solving for ω we obtain

$$\omega = \omega_c + \omega_Q \qquad (8.2)$$

where

$$\omega_c = \frac{\frac{\partial T}{\partial t} + \mathbb{V}_p \cdot \nabla_p T}{\frac{R}{c_p}\frac{T}{p} - \frac{\partial T}{\partial p}} \qquad (8.3)$$

and

$$\omega_Q = \frac{-\frac{1}{c_p}\frac{dQ}{dt}}{\frac{RT}{c_p p} - \frac{\partial T}{\partial p}} \qquad (8.4)$$

In Chapter IV we have given our method for calculating ω_c, the adiabatic part of the vertical motion. Daily evaluation of ω_Q is not possible at the present time because of the lack of instruments which are capable of measuring $\frac{dQ}{dt}$ in the stratosphere. Estimates of the mean heating rates have been made by Ohring (1958), Murgatroyd and Goody (1958)

and others. However, these estimates are not sufficient to compute on a daily basis and, as we shall show, they do not allow us to compute $[\overline{\omega_a}]$.

Using the continuity equation we find that the mean horizontal meridional velocities can be obtained from

$$[\overline{V}] = [\overline{V}]_0 \frac{\cos\phi_0}{\cos\phi} - \frac{[\overline{\omega_L}]-[\overline{\omega_U}]}{p_L - p_U} - \frac{a(\sin\phi - \sin\phi_0)}{\cos\phi} \tag{8.5}$$

Starting at our northern most latitude, we let $[\overline{V}]_0 = 0$ and successively obtained values of $[\overline{V}]$ every 5° down to the equator. Thus we have that part of the actual $[\overline{V}]_s$ due to the adiabatic motion.

Figures 8.1 and 8.2 show the latitudinal distribution of $[\overline{\omega_c}]$ and the $[\overline{V}]_{50}$ s computed from the $[\overline{\omega_c}]$ s. The $[\overline{\omega_c}]$ values have purposely been plotted upside down to give the appearance of upward and downward motion past pressure surfaces in the stratosphere.

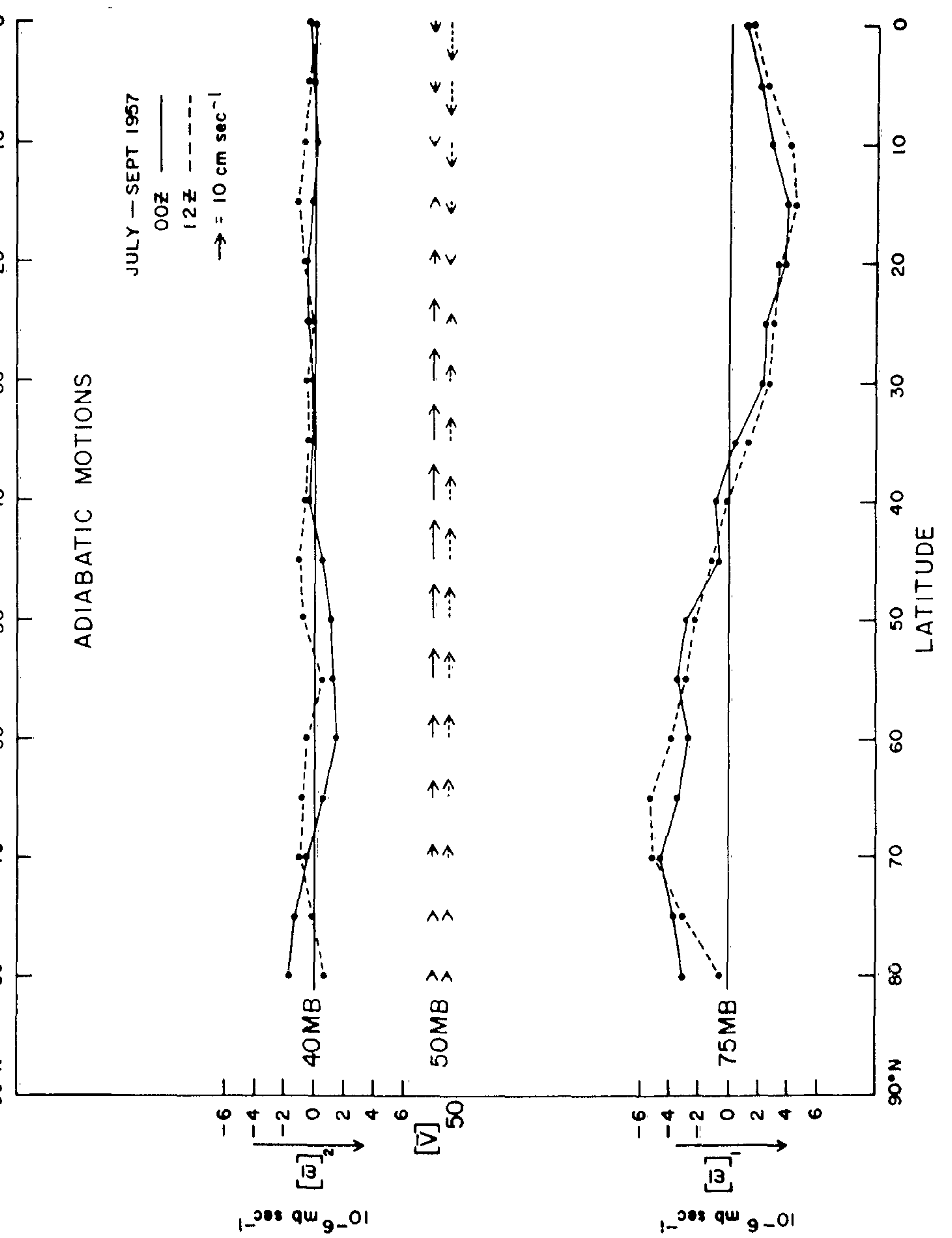
The individual $\overline{\omega_c}$ maps were drawn independently and Figures 8.1 and 8.2 show that, except at high latitudes for the period October through December at 75 mb, the 00Z and 12Z values are fairly close which lends some confidence to the computed values. The values of $[\overline{\omega_c}]$ at 75 mb for the period October through December were rechecked, and it was found that, using the $\overline{\omega_c}$ values, the analysis could not be altered sufficiently to give as good agreement between the October through December 00Z and 12Z values as between the 00Z and 12Z values at the other level and for the other three month period.

Summarizing, the mean motions obtained by this method under the adiabatic assumption are easily computed, and the similarity of the 00Z and 12Z values indicates that the signal has been separated from the noise by using three month averages. The omission of the diabatic motion is the greatest drawback, and, as we shall show next, the diabatic motions are as important as the adiabatic motions in the determination of the actual mean meridional motions.

Table 8.1 list the three month mean horizontal meridional velocities obtained from the observed winds. It should be noticed that the maximum average velocities are over one half meter per second. Such large mean horizontal meridional motions over a three month period are extremely difficult to reconcile with the observed changes in the angular momentum distribution. Also notice should be made of the fact that the agreement between the 00Z and 12Z values is very poor indicating that not enough data have been used to separate the signal from the noise.

The last column of table 8.1 lists the six month average of the observed horizontal meridional velocities using both the 00Z and 12Z data. The comparison with Jensen's (1960) circulation for January 1958 is good north of 30°N, but our data show that the indirect cell extends to 30°N with the direct cell further to the south for the period July through December 1957. Jensen's data extended just from 20°N to the North Pole. Thus it appears that we have obtained the major features of the horizontal meridional circulation at 50 mb by reducing six months of actual wind data.

Table 8.2 lists the mean horizontal meridional velocities obtained from the computed adiabatic vertical velocities by use of the



ADIABATIC MOTIONS
JULY — SEPT 1957
00Z
12Z
→ = 10 cm sec⁻¹
40MB
50MB
75MB
LATITUDE
90°N 80 70 60 50 40 30 20 10 0
[ω̄]₂ 10⁻⁶ mb sec⁻¹
[V̄]₅₀
[ω̄]₁ 10⁻⁶ mb sec⁻¹
-6 -4 -2 0 2 4 6

FIGURE 8.1

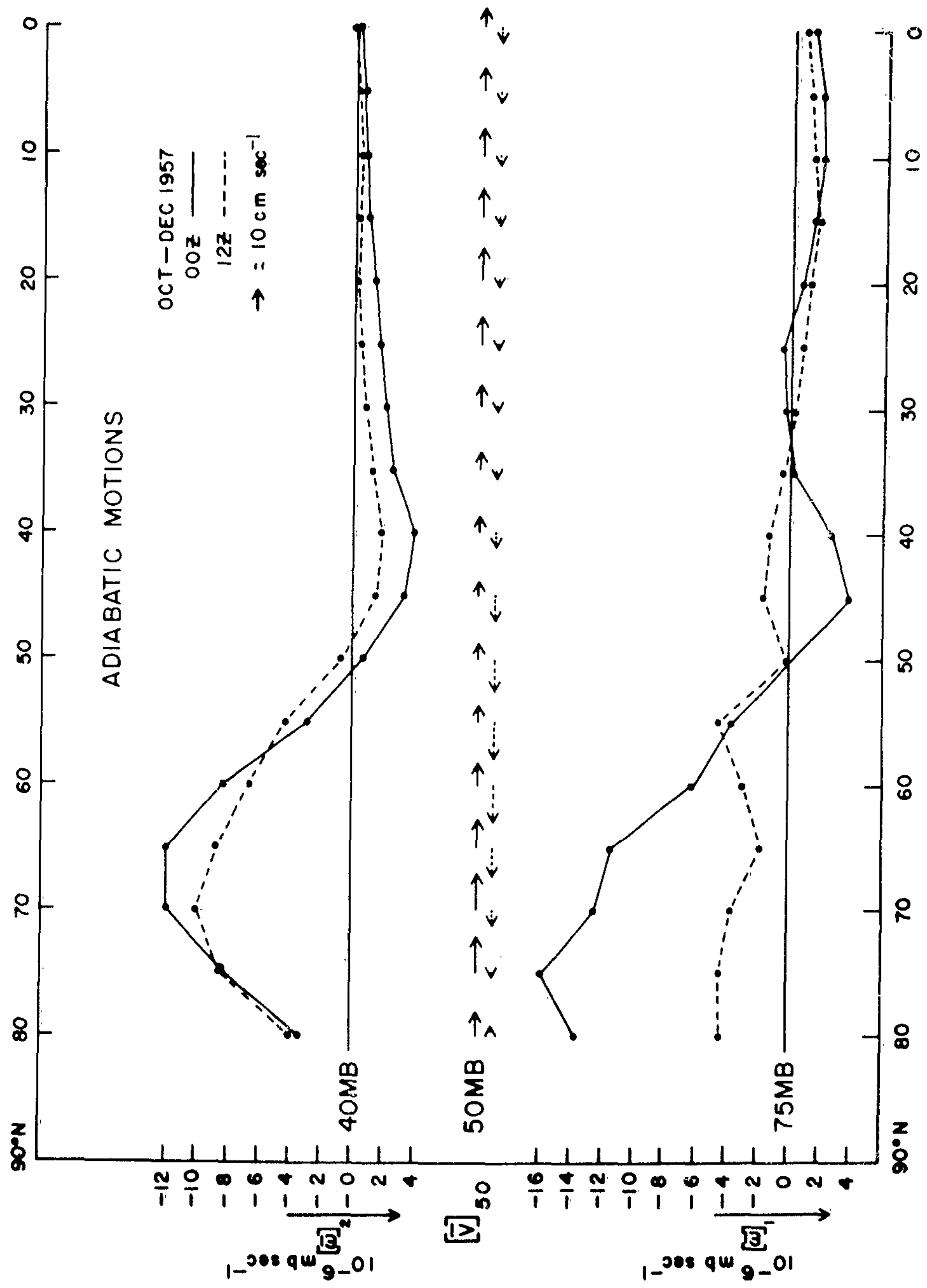

ADIABATIC MOTIONS
OCT—DEC 1957
00Z
12Z
= 10 cm sec⁻¹
40MB
50MB
75MB
90°N
LATITUDE
10⁻⁶ mb sec⁻¹
10⁻⁶ mb sec⁻¹

FIGURE 8.2

Table 8.1

Mean 50 mb meridional winds computed from the actual wind data. Units cm per second. Positive values indicate northward motion.

	July - September		October - December		Six month average
Lat.	00Z	12Z	00Z	12Z	
80°N	64	34	-21	19	24
	48	21	13	12	23
70	35	11	49	21	29
	3	1	48	21	18
60	-9	-7	18	-12	-3
	-1	0	-34	-5	-10
50	-0	7	-60	-8	-15
	-9	-2	-39	-31	-20
40	-14	0	-29	-39	-21
	-21	4	-3	-30	-13
30	-11	21	-3	-17	-3
	-5	29	-12	10	5
20	7	26	-17	22	9
	23	23	-14	18	13
10	28	19	-9	4	10
	18	9	1	-8	5
0°	10	5	13	-13	4

Table 8.2

Mean 50 mb meridional motions computed from the adiabatic motions. Units cm per second. Positive values indicate northward motion.

	July - September		October - December		Six month average	Diabatic motions
Lat.	00Z	12Z	00Z	12Z		
$80^{\circ}N$	-2	-2	-16	-0	-5	29
	-4	-4	-23	2	-7	31
70	-8	-8	-23	9	-7	36
	-12	-13	-19	17	-7	25
60	-16	-16	-14	22	-6	4
	-21	-20	-11	22	-7	-2
50	-25	-21	-11	20	-9	-6
	-27	-21	-10	16	-10	-10
40	-26	-19	-10	9	-11	-9
	-24	-16	-12	5	-12	-1
30	-21	-11	-15	3	-11	8
	-16	-6	-18	3	-9	15
20	-10	0	-19	3	-6	16
	-3	8	-19	5	-2	15
10	3	16	-17	7	2	9
	7	22	-14	8	6	-1
0°	7	25	-12	9	7	-4

continuity equation. First, we notice that the maximum values are about 1/4 of a meter per second which is close to theoretical estimates of values of the mean horizontal meridional velocities in the stratosphere. Because of the accumulation of errors in the adiabatically computed $[\bar{V}]$s due to the integration of the $[\bar{\omega}_c]$ values, we should direct our attention to the shape of the curves of the meridional distribution of the $[\bar{V}]$ as well as the actual magnitudes. The July through September period shows divergence at 50 mb north of roughly 45°N with convergence to the south for both 00Z and 12Z. The early winter period shows for both 00Z and 12Z convergence north of 55°N and south of 25°N with divergence in between.

The last column of table 8.2 gives the six month average horizontal meridional velocities at 50 mb due to diabatic effects. Equation (8.2) and the continuity equation can be used to show that $[\bar{V}]_Q = [\bar{V}] - [\bar{V}]_c$ where $[\bar{V}]$ is the mean horizontal meridional component of the actual wind, $[\bar{V}]_c$ is the mean horizontal meridional motion obtained from the mean vertical meridional adiabatic motions by the continuity equation, and the $[\bar{V}]_Q$ is the mean horizontal meridional motion obtained from the mean vertical meridional diabatic motions and the continuity equation. We notice that these mean horizontal meridional diabatic motions are roughly in the same direction as those obtained by Murgatroyd and Singleton (1961), but are much smaller in magnitude. Also we notice that the diabatic and adiabatic motions are of comparable size but often have the opposite directions.

We conclude from this that, in the mean, the diabatic heating is very important in the lower stratosphere and cannot be disregarded. This result is not surprising when we remember that the atmosphere is

a heat engine which is run by radiation from the sun, and therefore we should expect the diabatic effects to be detectable in the mean motions.

Before continuing, a few remarks should be made about $\frac{dQ}{dt}$, total radiation absorbed minus the total radiation emitted. The use of the total derivative sign is deceiving since the partial derivatives in the expansion of the total derivatives have no physical meaning. The symbol $\frac{dQ}{dt}$ should be replaced by a single symbol to avoid the misleading connotations. However, the symbol is well entrenched in the literature so there is not much hope for the suggested change.

Recently Murgatroyd and Singleton (1961) attempted to evaluate $[\bar{\omega}]$ using computed heating rates and observed temperature distributions. They use a relaxation method employing the continuity equation and only the mean terms from equation (8.1). They assume that the eddy terms may be disregarded.

Rewritting (8.1) and taking the time and latitude means gives

$$
\begin{aligned}
\frac{1}{c_p}\left[\overline{\frac{dQ}{dt}}\right] &= \frac{\partial[\bar{T}]}{\partial t} + \left[\overline{V_p \cdot \nabla_p T}\right] - \left[\overline{\omega \frac{RT}{c_p p}}\right] + \left[\overline{\omega \frac{\partial T}{\partial p}}\right] \\
&= \frac{\partial[\bar{T}]}{\partial t} + [\bar{v}]\frac{\partial[\bar{T}]}{\partial y} + \left[\overline{v' \frac{\partial T'}{\partial y}}\right] + [\bar{u}]\left[\overline{\frac{\partial T}{\partial x}}\right] + \left[\overline{u' \frac{\partial T'}{\partial x}}\right] \\
&\quad - [\bar{\omega}]\frac{R[\bar{T}]}{c_p p} + [\bar{\omega}]\frac{\partial[\bar{T}]}{\partial p} - \left[\overline{\omega' \frac{RT'}{c_p p}}\right] + \left[\overline{\omega' \frac{\partial T'}{\partial p}}\right] \qquad (8.6) \\
&\quad + \left[\bar{v}^* \frac{\partial \bar{T}^*}{\partial y}\right] + \left[\bar{u}^* \frac{\partial \bar{T}^*}{\partial y}\right] - \left[\bar{\omega}^* \frac{R\bar{T}^*}{c_p p}\right] + \left[\bar{\omega}^* \frac{\partial \bar{T}^*}{\partial p}\right]
\end{aligned}
$$

Since the term $\left[\dfrac{\partial \overline{T}}{\partial x}\right] = 0$

$$\frac{1}{c_p}\left[\overline{\frac{dQ}{dt}}\right] = \frac{\partial[\overline{T}]}{\partial t} + [\overline{V}]\frac{\partial[\overline{T}]}{\partial y} - [\overline{\omega}]\frac{R[\overline{T}]}{c_p p} + [\overline{\omega}]\frac{\partial[\overline{T}]}{\partial p}$$
$$+\left[\overline{V'\frac{\partial T'}{\partial y}}\right] + \left[\overline{u'\frac{\partial T'}{\partial x}}\right] - \left[\overline{\omega'\frac{RT'}{c_p p}}\right] + \left[\overline{\omega'\frac{\partial T'}{\partial p}}\right] \qquad (8.7)$$
$$+\left[\overline{V}^*\frac{\partial \overline{T}^*}{\partial y}\right] + \left[\overline{u}^*\frac{\partial \overline{T}^*}{\partial x}\right] - \left[\overline{\omega}^*\frac{R\overline{T}^*}{c_p p}\right] + \left[\overline{\omega}^*\frac{\partial \overline{T}^*}{\partial p}\right]$$

Using calculated heating rates for $\frac{1}{c_p}\left[\overline{\frac{dQ}{dt}}\right]$, Murgatroyd and Singleton used only the first four terms on the RHS of equation (8.7) plus the continuity equation to compute the meridional motion. White (1954) found that there is a countergradient heat flux in the lower stratosphere and we have reconfirmed his results. This flux of heat cannot be disregarded, and therefore Murgatroyd and Singleton were not justified in throwing out the terms of eddy heat flux. Murgatroyd and Singleton have tried to force a Hadley regime on the atmosphere where the evidence points to a combination of Hadley regime and a Rossby regime. Also their mean velocities seem to be too large to satisfy the angular momentum balance.

If it were possible to compute the mean diabatic vertical velocities then we could obtain the actual mean meridional circulation from

$$[\overline{\omega}] = [\overline{\omega}_Q] + [\overline{\omega}_c]$$

since we already have the $\left[\omega_c\right]$.

From (8.4) we obtain

$$\left[\overline{\omega_Q}\right] = -\frac{1}{c_p}\left[\overline{\frac{dQ}{dt}}\right]\left[\overline{\frac{1}{\frac{RT}{c_p p}-\frac{\partial T}{\partial p}}}\right] - \frac{1}{c_p}\left[\overline{\frac{dQ}{dt}}^{*}\;\overline{\frac{1}{\frac{RT}{c_p p}-\frac{\partial T}{\partial p}}}^{*}\right] - \frac{1}{c_p}\left[\overline{\frac{dQ}{dt}'\;\frac{1}{\frac{RT}{c_p p}-\frac{\partial T}{\partial p}}'}\right] \qquad (8.8)$$

The terms in $\dfrac{1}{\frac{RT}{c_p p}-\frac{\partial T}{\partial p}}$ may be computed from our data, but only mean values of the radiation and estimates of the time and space variations are available. Because of the variations in water vapor, ozone and carbon dioxide in the stratosphere and the change in the elevation of the sun in polar regions, we may assume that large absolute values of $\frac{dQ}{dt}'$ at a given station are an order of magnitude larger than large absolute values of $\left[\overline{\frac{dQ}{dt}}\right]$. We also assume that large absolute values of $\overline{\frac{dQ}{dt}}^{*}$ are at most an order of magnitude larger than large absolute values of $\left[\overline{\frac{dQ}{dt}}\right]$.

In the stratosphere $\frac{\partial T}{\partial p}$ is small compared to $\frac{RT}{c_p p}$ so, for the mean term,

$$\left[\overline{\frac{1}{\frac{RT}{c_p p}-\frac{\partial T}{\partial p}}}\right] = \frac{p\,c_p}{R\left[\overline{T}\right]} = \frac{3.44\,p}{\left[\overline{T}\right]} \qquad (8.9)$$

For our data we find that $\frac{\partial T'}{\partial p}$ is the same order of magnitude as $\frac{\partial T'}{p}$ so

$$\frac{1}{\frac{RT}{c_p p}-\frac{\partial T}{\partial p}} = \frac{1}{\frac{R\overline{T}}{c_p p}-\frac{\partial \overline{T}}{\partial p}+\frac{RT'}{c_p p}-\frac{\partial T'}{\partial p}} = \frac{1}{\frac{R\overline{T}}{c_p p}+\frac{RT'}{c_p p}-\frac{\partial T'}{p}}. \qquad (8.10)$$

Hence

$$\overline{\frac{1}{\frac{R}{c_p}\frac{T}{p} - \frac{\partial T}{\partial p}}}^{\,\prime} \doteq \frac{\frac{T'}{p}\left(2 - \frac{R}{c_p}\right)}{\left(\frac{R}{c_p}\frac{\overline{T}}{p}\right)^2} \doteq \frac{T'}{\overline{T}^2}\, 20\, p' \qquad (8.11)$$

disregarding third order and higher terms, this being permissible since

$$\frac{T'}{\overline{T}} \doteq \frac{5^\circ}{200^\circ} << 1$$

From (8.10)

$$\overline{\frac{1}{\frac{RT}{c_p p} - \frac{\partial T}{\partial p}}} \doteq \frac{\frac{c_p}{R}p}{\overline{T}} = \frac{\frac{c_p}{R}p}{[\overline{T}] + \overline{T}^*} \qquad (8.12)$$

Thus

$$\overline{\frac{1}{\frac{RT}{c_p p} - \frac{\partial T}{\partial p}}}^{\,*} \doteq \frac{-\frac{c_p}{R}p\overline{T}^*}{([\overline{T}] + \overline{T}^*)^2} \doteq \frac{-\frac{c_p}{R}p\overline{T}^*}{[\overline{T}]^2} = -\frac{\overline{T}^*}{[\overline{T}]^2}\, 3.44\, p$$

Equation (8.8) now can be written

$$[\overline{\omega}_Q] = -\frac{1}{c_p}\left[\overline{\frac{dQ}{dt}}\right]\frac{3.44\,p}{[\overline{T}]} + \frac{1}{c_p}\left[\overline{\frac{dQ}{dt}}^{\,*}\frac{\overline{T}^*}{[\overline{T}]^2}\, 3.44\, p\right]$$

$$- \frac{1}{c_p}\left[\overline{\frac{dQ}{dt}'\frac{T'}{\overline{T}^2}\, 20\, p}\right] \qquad (8.13)$$

Using $T' \doteq \overline{T}^* \doteq 5^\circ$, $\overline{T} \doteq [\overline{T}] \doteq 200^\circ$ taking maximum values (i.e., perfect correlation) of the last two terms of (8.13) we obtain

$$[\overline{\omega}_Q] = -\frac{1}{c_p}\left[\overline{\frac{dQ}{dt}}\right]\frac{3.44\,p}{[\overline{T}]}(1) + \frac{1}{c_p}\left[\overline{\frac{dQ}{dt}}\right]\frac{3.44\,p}{[\overline{T}]}(0.25)$$

$$- \frac{1}{c_p}\left[\overline{\frac{dQ}{dt}}\right]\frac{3.44\,p}{[\overline{T}]}(1.20) \qquad (8.14)$$

indicating that the eddy terms cannot be disregarded unless the correlations are very small or zero.

From Murgatroyd and Singleton we obtain a maximum value of $\frac{1}{c_p}\left[\overline{\frac{dQ}{dt}}\right]$ equivalent to a heating rate of $1^{\circ}K$ per day at 75 mb. This gives a value of 15 $\times 10^{-6}$ millibars per second for the first term on the RHS of equation (8.13). This is slightly larger than the magnitudes of $\left[\overline{\omega_c}\right]$. On the other hand table 8.2 indicates that the $\left[\overline{\omega_Q}\right]$ s are the same magnitude as the $\left[\overline{\omega_c}\right]$ s. Thus either the values of $\frac{1}{c_p}\left[\overline{\frac{dQ}{dt}}\right]$ from Murgatroyd and Singleton are too large or the first term on the RHS of (8.8) is balanced by the transient eddy term.

Figures 8.1 and 8.2 show that, for adiabatic motions, we have rising motion in the polar regions and sinking in equatorial regions. The indications from table 8.2 are that the mean meridional diabatic vertical motions and the mean meridional adiabatic vertical motions tend largely to oppose each other, leaving the true mean meridional vertical motions. Thus we would expect on these grounds to find mean diabatic sinking motion in polar regions from July through December 1957.

Returning to equation (8.8) we may obtain the equation

$$\left\{\left[\overline{\omega_Q}\right]\right\} = -\frac{1}{c_p}\left\{\left[\overline{\frac{dQ}{dt}}\right]\right\}\left\{\left[\overline{\frac{1}{\frac{RT}{c_p p}-\frac{\partial T}{\partial p}}}\right]\right\} - \frac{1}{c_p}\left\{\left[\overline{\frac{dQ}{dt}}\right]''\left[\overline{\frac{1}{\frac{RT}{c_p p}-\frac{\partial T}{\partial p}}}\right]''\right\}$$

$$-\frac{1}{c_p}\left\{\left[\overline{\frac{dQ}{dt}}^{*}\,\overline{\frac{1}{\frac{RT}{c_p p}-\frac{\partial T}{\partial p}}}^{*}\right]\right\} - \frac{1}{c_p}\left\{\left[\overline{\frac{dQ}{dt}'\,\frac{1}{\frac{RT}{c_p p}-\frac{\partial T}{\partial p}}'}\right]\right\}$$

In the lower stratosphere we find that the term

$$\left[\overline{\frac{1}{\frac{RT}{c_p p}-\frac{\partial T}{\partial p}}}\right]$$

is almost constant from the pole to the equator. For long time periods the mean term $\left\{\left[\overline{\frac{dQ}{dt}}\right]\right\}$ has a magnitude of about one tenth of the large values of $\left[\frac{dQ}{dt}\right]$ in the lower stratosphere from Murgatroyd and Singleton's data. Hence we are left with the conclusion that the transient eddy term is the most important term for computing the mean diabatic vertical motion over the northern hemisphere for a long time period in the lower stratosphere.

For our purposes, we conclude that the adiabatic vertical motion alone cannot replace the actual vertical motion in evaluation of meridional eddy terms and mean terms of our integrals containing ω.

Employing equation (8.4) and an instantaneous, large value of $\frac{1}{c_p}\frac{dQ}{dt}$ corresponding to 5°K per day we obtain 5×10^{-5} millibars per second as a large value of ω_Q in the lower stratosphere. Large values of ω_c were about 8×10^{-4}. Thus, unless the correlation of the diabatic vertical motion with the other quantities is especially high, we may use the adiabatic vertical motions for the actual vertical motions to obtain representative values of transient eddy terms of integrals containing ω. Even though the magnitude of the large time averages of ω_c is smaller than the magnitude of the large individual ω_c, it is still larger than the magnitude of large $\overline{\omega_Q}$ so we have used the adiabatic vertical motions to evaluate the standing zonal eddy terms in ω.

Summary

The 50 mb values of $[\overline{V}]$ computed by taking time and latitudinal means of the observed south to north component of the wind were found to be unreliable for three month periods but seem to be significant for the six month period.

An order of magnitude study shows that large magnitude individual values of ω_c and of $\overline{\omega_c}$ are greaterthan large magnitude values of ω_Q and $\overline{\omega_Q}$ respectively, but values of $[\overline{\omega_c}]$ and $[\overline{\omega_Q}]$ have comparable magnitudes. Under the assumption that the time and zonal correlations of ω_Q with the height, temperature and wind are not as large as the correlations of ω_c with these quantities, we take the time covariances and zonal covariances with ω_c to be representative of the covariances with ω, the actual vertical velocities.

The diabatic effects are not easily determined, but the time variations of $\frac{dQ}{dt}$ are as important as the meridional variations of $\left[\frac{dQ}{dt}\right]$ in determing the meridional circulation. Residual motions show that the meridional circulations due to diabatic effects are of the same magnitude as adiabatic effect and cannot be disregarded Hence meridional terms calculated using the mean adiabatic vertical motion are not considered as representative of the atmosphere at these levels.

IX Approximation of the Quantities $\frac{p}{g}\iint \frac{\partial \psi}{\partial t}\,dx\,dy$, $\frac{\partial}{\partial t}\iiint (\varphi + J)\frac{dp}{g}\,dx\,dy$ and $\frac{\partial}{\partial t}\iiint E\,\frac{dp}{g}\,dx\,dy$

The integral

$$\frac{p}{g}\iint \frac{\partial \psi}{\partial t}\,dx\,dy \qquad (9.1)$$

has been evaluated at three levels; 100 mb, 50 mb, and 30 mb using maps published by the Weather Bureau. Figure 7.1 in Chapter VII shows the mean heights for the latitudes at the beginning and end of the two periods. These values were used to evaluate this integral. Table 9.1 gives the values for the integral at the three levels for each period. The units are ergs cm^{-2} sec^{-1}. Since this integral is a boundary integral which applies both at the top and the bottom of the layer considered, the difference between the top and bottom values must be used in the evaluation of the change of the potential plus internal energy, equation (9.2). The last three columns of Table 9.1 give the values which are used for evaluating equation (2.14) for the three layers. indicated.

The integral (9.1) represents the change of potential energy of the layer considered due to the movement of the layer as a whole. It does not include changes of potential energy due to the internal redistribution of mass within the layer. Negative values indicate a lowering of the pressure surfaces. As one would expect the potential

Table 9.1

Values of $\frac{p}{g}\iint \frac{\partial\psi}{\partial t}\, dx\, dy$ at three levels, and contribution of $\frac{p}{g}\iint \frac{\partial\psi}{\partial t}\, dx\, dy \Big|_{p_u}^{p_L}$ to the time rate of change of potential plus internal energy for 3 layers over the northern hemisphere. Units are ergs cm^{-2} sec^{-1}.

	July - September	Oct. - December 1957
Level		
100 mb	-177	-330
50 mb	-112	-201
30 mb	-76	-148
Layer		
100 mb - 50 mb	-65	-129
50 mb - 30 mb	-36	-53
100 - 30 mb	-101	-182

Table 9.2

Values of the potential and internal energy integrals. Units are ergs cm^{-2} sec^{-1}.

	July-September 1957			October - December 1957		
	100 - 50 mb	50 - 30 mb	100 - 30 mb	100 - 50 mb	50 - 30 mb	100 - 30 mb
$\frac{\partial}{\partial t}\iiint \phi \frac{dp}{g}\,dx\,dy$	-100	-48	-148	-183	-90	-273
$\frac{\partial}{\partial t}\iiint I \frac{dp}{g}\,dx\,dy$	-118	-51	-169	-217	-125	-342
$\frac{\partial}{\partial t}\iiint (\phi + I) \frac{dp}{g}\,dx\,dy$	-218	-99	-317	-400	-215	-615

energy of the layer decreases during both periods as the upper and lower pressure surfaces move downward.

Using the above information plus the information on the rate of change of temperature for the two periods, obtained from the Weather Bureau maps (see figures 7.2, 7.3 and 7.4), it was possible to compute values for the LHS of equation (2.14),

$$\frac{\partial}{\partial t}\iiint(\mathcal{P}+\mathcal{I})\frac{dp}{g}\,dx\,dy. \tag{9.2}$$

Table 9.2 gives the rate of change of potential energy, the rate of change of internal energy and the rate of change of potential plus internal energy for the layers and periods indicated. The integrals were evaluated by using linear interpolation between the levels of observation.

The values for the actual rate of change of kinetic energy during the periods,

$$\frac{\partial}{\partial t}\iiint E\,\frac{dp}{g}\,dx\,dy, \tag{9.3}$$

are given in Table 9.3 (see equation 2.9). These values were obtained by computing the geostrophic wind components from the grid-point height values taken from the daily Weather Bureau maps. Since the geostrophic assumption is not valid in the equatorial regions, only the area from 10°N to 85°N was to be used to evaluate this integral. The values in ergs

cm^{-2} sec^{-1} for this area were taken as the values of (9.2). It is realized that there may be very strong winds in the equatorial regions which may be very important in determining the kinetic energy over the northern hemisphere. The Weather Bureau maps went down to only $5^{o}N$ and it was difficult to ascertain the kinetic energy in the inner tropical region using the observed winds.

The Weather Bureau maps in these higher levels depend heavily on continuity, particularly because of the small number of observations. It was decided not to plot daily maps to obtain the wind speeds to evaluate this integral. Because of the sparsity of observations, 3 to 5 day mean maps would be necessary to obtain any sort of picture of the equatorial flow. The clustering of stations over land masses would decrease the reliability of the values obtained from such maps.

The decrease in kinetic energy from July to September is also indicated by the flattening of the height curves from the first of July to the end of September, Figure 7.1. These figures indicate an increase in kinetic energy for the period October-December which is noted in Table 9.3

Using the values of $\left\{ \left[\overline{C^2} \right] \right\}$ computations were made which gave close agreement with the values in Table 9.3. It is realized that the day to day fluctuations of kinetic energy as computed from daily maps are large, hence values of (9.2) are subject to error due to our method of computation.

Table 9.3

Values of $\frac{\partial}{\partial t}\iiint E \frac{dp}{g}\, dx\, dy$ for three layers.
Units are ergs cm^{-2} sec^{-1}.

	July - September	October - December 1957
100 - 50 mb	-.36	+13.93
50 - 30 mb	-1.14	+4.62
100 - 30 mb	-1.50	+18.55

X Approximations of the Quantities $\iiint \omega\alpha \frac{dp}{g} dx\,dy$ and $\iint \frac{c_p \omega T}{g} dx\,dy$.

As previously described each integral evaluated is separated into three or four terms. The first term is the transient eddy effect, and the second the zonal standing eddy effect. Values of these two terms are given for all of the approximated integrals. Values of the meridional standing eddy term and the mean term are given for some of the approximated integrals, but are not given for other integrals where the values were felt to be unrepresentative. Where applicable, reasons for exclusion of these latter two terms are included in the discussion of the individual integral. Values of the pressure eddy term were not calculated since, at most, only a few pressure levels were available for the volume integrals. Vertical boundary terms at the equator do not contain meridional eddies, and "horizontal" boundary terms do not contain pressure eddy terms.

The integrals

$$\iiint \omega\alpha \frac{dp}{g} dx\,dy = \iint \omega T \frac{R}{g} \frac{dp}{p} dx\,dy \qquad (10.1)$$

and

$$\iint \omega T \frac{c_p}{g} dx\,dy \qquad (10.2)$$

will be discussed together since they are functions of the surface integral

$$\iint \omega T\, dx\,dy. \qquad (10.3)$$

As shown in Chapter IV, the vertical motion can be divided into an adiabatic part and a diabatic part. Only the adiabatic part can be calculated from the observations. Using estimations of the diabatic vertical motion derived from available diabatic heating rates (Murgatroyd & Singleton, 1961) we have concluded that the adiabatic vertical motion cannot be used to determine the mean motion or the meridional motion over the northern hemisphere. Hence values of the mean terms and meridional eddies calculated using the adiabatic vertical motion are not considered to be representative of the actual atmosphere. Between 100 mb and 30 mb, below the region of ozone heating, we do believe that the individual, outstanding features of the adiabatic vertical motion field represent the individual, outstanding features of the true vertical motion field. For this reason we have taken the values of the transient and standing zonal eddies, calculated using the adiabatic vertical motions, as representatives of the actual transient and standing zonal eddies over the northern hemisphere.

The mean term of (10.3) presents an additional problem. The mean value of T is around 200°K while the departures of T used in calculating the eddies are of the order of 5°. The mean values and departures of ω_c have equivalent magnitudes. Considering the above and remembering that the correlation coefficient of ω and T in the eddy terms is less than one, we see that the mean terms are the largest contributors to the integrals (10.1) and (10.2) over the hemisphere.

If we had considered the whole sphere, then the mean term would be identically zero since

$$\oiint \omega \, dx\, dy = 0 \qquad (10.4)$$

for any pressure level in the stratosphere.

Murgatroyd and Singleton (1961) obtained meridional circulations in and above the stratosphere using net radiation values. Their lower stratospheric motions are almost the opposite of our adiabatic motions, but the magnitudes are about the same. Since both meridional models do not seem to fit with the required angular momentum balance, it is apparent that the true picture lies some place in between.

The adiabatic term

$$\iiint \omega_e \, \alpha \, \frac{dp}{g} \, dx \, dy \tag{10.5}$$

can be considered as a transformation term between kinetic energy and potential plus internal energy due to adiabatic processes. However we must keep in mind that the internal energy is being constantly changed by diabatic processes in the actual atmosphere. Hence this is only the adiabatic redistribution of energy between the two forms.

The diabatic term

$$\iiint \omega_Q \, \alpha \, \frac{dp}{g} \, dx \, dy \tag{10.6}$$

measures a redistribution of energy between the two forms due to diabatic processes. Further discussions of diabatic terms are contained in Chapter XVI

Maps not reproduced here provided the values for the transient eddy part of the integrals. The similarities in the gross details of the 00Z and 12Z maps are reflected in the values of the transient eddies as seen in tables (10.1) and (10.2) where the upper values are for 00Z and the lower for 12Z. The values for the layer from 50 mb to 30 mb (40 mb level) are small and the change of sign from 00Z to 12Z indicates that the true value is probably not significantly different from zero.

In the average from 100 mb to 50 mb the adiabatic transient eddies and zonal eddies seem to convert kinetic energy into potential plus internal, while from 50 mb to 10 mb the transfer seems to be in the opposite direction for both periods. The magnitudes of these values and values of $\frac{\partial}{\partial t}\iiint E \frac{dp}{g}\, dx\, dy$ (Table 9.3) indicates that the eddy terms of (10.1) cannot be disregarded in the equation for the rate of change of total kinetic energy.

On the other hand the values for $\frac{\partial}{\partial t}\iiint (\theta + J)\frac{dp}{g}\, dx\, dy$ (Table 9.2) indicates that the transient eddy and zonal eddy terms in (10.1) and (10.2) (given in Tables 10.1 and 10.2) are only minor consideration in the equation for the time rate of change of potential plus internal energy.

The meridional eddy values and the mean term values are given but these represent only the adiabatic part and we feel that they do not represent the true values for the atmosphere.

Table 10.1

Values of $\iiint \frac{\omega_e \alpha}{g} dp\, dx\, dy$ using the adiabatic part of ω

Units are ergs cm^{-2} sec^{-1}.

		July-September 1957			October-December 1957		
		100 - 50 mb	50 - 30 mb	30 - 10 mb	100 - 50 mb	50 - 30 mb	30 - 10 mb
Transient	00Z	3.5	0.6	4.7	5.0	2.2	-0.3
Eddies	12Z	7.3	-1.9		4.4	2.4	
Zonal	00Z	3.8	0.2	-5.7	9.0	-2.6	-139.0
Eddies	12Z	2.2	-0.7		6.2	-4.0	
Meridional	00Z	-41.4	1.2	-5.7	-26.6	3.8	-24.0
Eddies	12Z	-46.2	0.4		-19.2	1.8	
Mean	00Z	240.0	10.0	735.6	-371.0	-98.0	918.6
Term	12Z	371.0	-175.0		-174.0	-293.0	

Table 10.2

Values of $\iint \frac{c_p T \omega}{g} \, dx\,dy$ using the adiabatic part of ω.

Units are ergs cm^{-2} sec^{-1}.

		July- September 1957			October-December 1957		
		75 mb	40 mb	20 mb	75 mb	45 mb	20 mb
Transient	00Z	18	4	15	26	15	-1
Eddies	12Z	37	-13		22	16	
Zonal	00Z	19	1	-18	45	-18	-440
Eddies	12Z	11	-5		31	-27	
Meridional	00Z	-207	8	-18	-133	+26	-76
Eddies	12Z	-231	+3		-96	12	
Mean	00Z	1202	+71	2330	-1853	-664	+2910
Term	12Z	1857	-1190		-870	-1992	

XI <u>Approximation of</u> $\iint \frac{\omega E}{g} dx dy$

Values of the surface integral

$$\iint \frac{\omega_c E}{g} dx dy \tag{11.1}$$

using the adiabatic ω_c were computed, and, as previously found by Roberts (1960), Jensen (1960) and others for the tropospheric regions, the vertical advection of kinetic energy is an order of magnitude smaller than the vertical advection of potential energy.

We assumed that values of the integral

$$\iint \frac{\omega_Q E}{g} dx dy, \tag{11.2}$$

the advection of kinetic energy due to the diabatic part of the vertical motion, were no larger in magnitude than the values of the integral (11.1). If Charney's (1948) scale analysis is valid for the longer, planetary waves found in the stratosphere, then we would expect that the sum of the values of (11.1) and (11.2) would be an order of magnitude smaller than the integral

$$\iint \frac{\omega \varphi}{g} dx dy \tag{11.3}$$

which is also found in equation (2.9), the rate of change of the kinetic

energy. Thus the vertical advection of kinetic energy can be disregarded in equation (2.9) since the values are smaller then the values for both (11.3) and (9.3).

The integral (11.3) is discussed in Chapter XV. The mean term of (11.3) is zero when the integration is over the sphere since

$$\iint \omega \, dx \, dy \equiv 0$$

when the integration is over the sphere.

Even though there was very little similarity between 00Z and 12Z maps of $\overline{\omega' c^{2\prime}}$ for the same level and same period, there was one very interesting feature which showed up upon analyzing the maps. On all but two of these maps there were centers of strong covariance. These centers were in most cases due to an individual station, the implication being that there were data errors. The data had already been well screened and no obvious errors could be found. Nolan and White (1959) and others obtained, but disregarded, such values. We found too many to disregard.

Two possible causes are suggested, but, since the integral (11.1) turned out to be not important in this study neither possible cause was investigated thoroughly in the present study.

The first possible cause could be due to the fact that the wind speed figures predominately in computing ω_c (see Chapter IV) . Thus $\overline{\omega_c' c^{2\prime}}$ might be expected to be very sensitive to errors when the wind speed is very large.

The second possible cause was suggested to the author by a standard hydrodynamical experiment in a density channel. If slightly denser colored water is allowed to flow under clear, lighter water along a sloping bottom of a vessel as shown in Figure 11.1

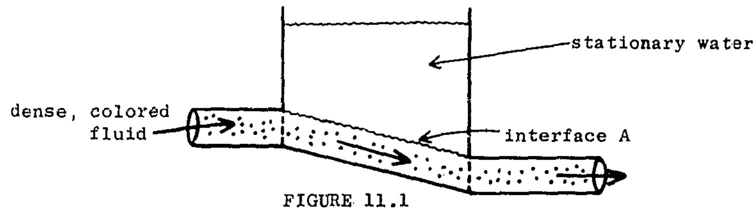


FIGURE 11.1

it is found that, above a critical shear across the interface A, instabilities occur along the interface causing plumes of colored water to rise up into the clear water. The strong centers of covariance on the $\overline{\omega' c'}$ maps have both negative and positive signs indicating strong transient eddy transports of kinetic energy in both directions. Hence we suggest the possibility that we have occasionally taken measurements in such plumes occuring in the stratosphere. Vertical velocities of over 12 ft. per second in the stratosphere as recorded by U-2 aircraft (NASA, 1958) lend credence to this theory. However, a study of shearing instability in the stratosphere should be made and then applied to the individual cases. This is left as a topic for further research since the results are of minor importance to this study.

Table 11.1

Values of $\iint \frac{\omega E}{g}\, dx\, dy$ using the adiabatic part of ω. Units are ergs cm $^{-2}$ sec $^{-1}$.

	July - September			October - December 1957		
	75 mb	40 mb	20 mb	75 mb	40 mb	20 mb
Transient	.47	.04	-.61	-.49	.28	1.88
Eddies	.01	-.03		-1.20	1.90	
Zonal	-.24	-.06	-1.06	.45	-.73	-.71
Eddies	.15	.15		-2.18	-.99	
Meridional	.80	-.18	.64	-.28	-2.11	-.70
Eddies	.85	-.18		-.49	-1.53	
Mean	.44	-.01	1.39	-1.32	-.46	-.39
Term	.59	-.26		-1.16	-.79	

XII <u>Approximation of the Quantity.</u> $\iint VE \frac{dp}{g} dx$

This term appears as a boundary integral in equation (2.9) and represents the flux of horizontal kinetic energy across the equatorial boundary into the considered mass.

$$\iint VE \frac{dp}{g} dx \qquad (12.1)$$

The zonal standing eddies were evaluated from the mean maps of $\overline{V}$ and $\overline{C^2}$. Values for the mean term of (12.1) were obtained by using the equatorial $[\overline{V}]$ values (computed from the adiabatic $\{[\overline{\omega}_c]\}$ and the continuity equation) and the equatorial values of $[\overline{E}]$ obtained from the $\overline{C^2}$ maps.

Table 12.1 gives our results for the transient eddy, zonal standing eddy and mean term components of (12.1).

We notice from Table 12.1 that the zonal eddy contributions were an order of magnitude smaller than the transient eddy contributions, and that the transient eddy values changed sign in three of the four cases. This suggest that there was a large random element in the values of $\overline{V'C^{2\prime}}$ which indeed was noticed when analyzing these maps.

This large amount of noise was probably due to the fact that, since V is also a function of C, values of $\overline{V'C^{2\prime}}$ were extremely sensitive to errors in C. Since the fields of both $\overline{V}$ and $\overline{C^2}$ were smoothed by the averaging and analyzing processes, the zonal standing eddy values did not contain as much noise.

Even though these values are large compared to those for the rate of change of the total kinetic energy, as shown in Table 15.3, we believe that the integral (12.1) is of minor importance in equation (2.9) when applied to the stratosphere over the northern hemisphere. If the whole sphere were considered, this integral would vanish.

Table 12.1

Values of $\iint V E \frac{dp}{g} d\chi$ Units are ergs cm^{-2} sec^{-1}.

	July-September		October-December 1957		
	100-50 mb	50-39 mb	100-50 mb	50-30 mb	
Transient	3.4	1.6	-5.5	-3.4	00Z
Eddies	-4.3	.2	3.6	1.6	12Z
Zonal	.48	.06	-.04	.01	00Z
Eddies	-.12	-.16	.14	-.03	12Z
Mean	.7	.10	-.8	-.19	00Z
Term	1.3	-.13	-.2	-.16	12Z

XIII Approximation of $\iint \psi V \frac{d\varphi}{g} dx$ (13.1)

Because of differing types of instruments used, we would expect that there might be a larger variation in space than in time for the quantity ψ and indeed this is what was found. This in turn has an effect on the spacial correlation, giving spurious, large values for the zonal standing eddies of (13.1) as seen in Table 13.1.

The term (13.1) is found in the equation for the rate of change of kinetic energy. Following Starr's (1951, b) derivation and interpretation of the various terms in the x, y, z coordinate system, we may consider (13.1) as a work term representing the work done on the mass at the equatorial boundary.

The values for transient and standing eddies along the equator are given in Table 13.1. Even though the sum of the transient and zonal standing eddies is comparable to the rate of change of the total kinetic energy, it is felt that the large values of the zonal eddies are not representative and that the eddy terms of (13.1) may be disregarded when considering the rate of change of the total kinetic energy.

The mean terms as computed from the data are heavily dependent on the total magnitude of the height. Thus any errors in $[\bar{V}]$ at the equator are greatly magnified. Since we have very little confidence in the values of $[\bar{V}]$ at the equator, we have not given any values for the mean terms.

Since the $[\bar{V}]$s and the $[\bar{\omega}]$s are related through the continuity equation, we would expect to find a partial compensation of the mean term of (13.1) and the mean term of the integral

$$\iint \frac{\varphi\omega}{g}\,dx\,dy \qquad (13.2)$$

in the equation for the rate of change of kinetic energy.

From Table 13.1 we conclude that the eddy terms of (13.1) may be disregarded in evaluating the rate of change of the total kinetic energy of the stratosphere over the northern hemisphere. The mean terms cannot be accurately calculated, but should to some extent be compensated by the mean terms of (13.2).

When considering the whole sphere, the boundary term (13.1) would not appear and the mean term of (13.2) would be zero since, over the sphere,

$$\iint \omega\,dx\,dy \equiv 0.$$

Table 13.1

Values of $\iint \frac{\psi v}{g} \, dp \, dx$. Units are ergs cm^{-2} sec^{-1}.

		July - September		October - December 1957	
		100 - 50 mb	50 - 30 mb	100 - 50 mb	50 - 30 mb
Transient	00Z	.2	.0	.1	.1
Eddies	12Z	.0	.1	.0	.1
Zonal	00Z	-.2	.2	10.0	-3.0
Eddies	12Z	-.6	-2.0	3.8	.1

XIV Approximation of $\iint \frac{c_p T V}{g}\, dp\, dx$ (14.1)

This equatorial boundary integral was evaluated using the actual temperatures and V components of the wind as taken from the soundings. The integral represents the advection of potential plus internal energy across the equator and appears in the equation for the time rate of change of potential plus internal energy.

The curves of the latitude means of $\left[\overline{T'V'}\right]$ for 00Z and 12Z were quite similar indicating that the similarity between the 00Z and 12Z maps is real.

Values for the transient eddies and zonal standing eddies are given in Table 14.1.

Even though the values change from plus to minus between 00Z and 12Z at the equator, the magnitude of the eddy contribution is small compared to the time rate of change of the potential plus internal energy. Hence we conclude that the eddy terms of (14.1) evaluated at the equator may be disregarded in the equation for the rate of change of potential plus internal energy.

Values of the mean terms of (14.1) are not given because of the large errors in $[\overline{V}]$ at the equator which are exaggerated by the total value (rather than a deviation) of the temperature.

Also, reasoning similar to that used in Chapter XIII indicates that the mean term is mostly cancelled by the mean term of

$$\iint \frac{c_p T \omega}{g}\, dx\, dy \qquad (14.2)$$

If we had integrated over the whole sphere, the boundary integral (14.1) would not appear and the mean term of (14.2) would be zero since

$$\iint \omega \, dx \, dy \equiv 0 .$$

for the whole sphere at each pressure level.

Table 14.1.

Values of $\iint \frac{c_p T V}{g} \, dp \, dx$. Units are ergs cm^{-2} sec^{-1}.

	July - September			October - December 1957			
	100 -50	50-30	30-10	100-50	50-30	30-10	
Transient	-5.0	2.5	39.7	21.6	7.6	21.1	00Z
Eddies	+10.2	-0.6	35.1	-14.0	-14.3	12.5	12Z
Zonal	15.4	-5.0	-57	62.6	-9.9	-143	00Z
Eddies	-1.5	1.3	-52	9.1	-4.8	-139	12Z

XV Approximation of $\iint \frac{\psi\omega}{g}\,dx\,dy$ (15.1)

This integral appears as a boundary integral in the equation for the rate of change of kinetic energy. By deriving the equation for the rate of change of potential plus internal energy in a slightly different manner, it can be shown that the integral appears in that equation too (see Jensen (1960)). The physical interpretation of this term is still very much in question (Pfeffer (1957)). Jensen (1960) interprets it as a measure of part of the transformation between kinetic and potential plus internal energy. A parallel derivation of this term in the x, y, z coordinate system show that this is part of what Starr (1951,b) refers to as the term representing the work on the boundary of the volume under consideration. This term is not zero only because the boundary pressure surfaces are not everywhere parallel to the geopotential surfaces. Over the whole sphere for any pressure surface in the stratosphere the surface integral of ω is identically zero. Thus the mean term of (15.1) is zero over the sphere which means that any contributions of (15.1) are due to non-zero correlations of ω and $\frac{\psi}{g}$.

The only ω s available were the adiabatic ones. Investigation of the diabatic ω s indicated that the transient and zonal eddies in ω could be approximated by replacing ω by ω_c, the adiabatic part. However, this cannot be done for the meridional eddies and mean terms in ω due in part to the relative large dependence of ω_Q on latitude.

Values for the transient eddy and zonal eddy part of the integral (15.1) using the adiabatic vertical motion are given in Table 15.1.

Where two values are given the upper is for 00Z and the lower is for 12Z. The variations in the values are a reflection of the inaccuracies in measuring both ω_c and H . Unfortunately, (15.1) turns out to be one of the most important terms in equation (2.9) when we consider the magnitudes of the various terms.

There is very little similarity between the 00Z and 12Z maps of the transient eddy term except over the U.S. during the period July through September 1957 at 75 mb. This dissimilarity is thought to be caused by the large errors in the heights; the height being an integral of the temperature from the earth's surface to the pressure level considered. The fact that the individual ω_c values at 40 mb are generally smaller than at 75 mb causes the 40 mb covariance maps of $\overline{\omega' H'}$ to have smaller values than the 75 mb maps $\overline{\omega' H'}$. However, the coverage of large areas of one sign at 40 mb gives mean values of $\left\{ \left[\overline{\omega' H'} \right] \right\}$ as large as at 75 mb.

Summing the eddy contributions for both periods indicates that the eddies transfer potential energy upward across both the 75 mb and 40 mb levels. The values at 20 mb are very suspect, but point to a downward flux of potential energy. 20 mb values were obtained by the buckshot method because of the small number of consecutive observations reaching 10 mb.

Table 15.1

Values of $\iint \frac{\psi\omega}{g}\,dx\,dy$ using adiabatic vertical motions for ω .
Units are ergs cm^{-2} sec^{-1}.

	July-September 1957			October-December 1957		
	75 mb	40 mb	20 mb	75 mb	40 mb	20 mb
Transient	17.74	13.73	-2	-25.12	38.30	-19
Eddies	-1.26	-6.30		14.99	7.30	
Zonal	-.34	2.18	-6.10	-2.24	-20.12	-48
Eddies	4.23	-3.12		4.27	-17.80	

XVI The integrals $\iiint \frac{dQ}{dt} \frac{dp}{g} dx dy$ and $\iiint \nabla_p \cdot \mathbb{F} \frac{dp}{g} dx dy$

Using the adiabatic vertical motions to evaluate the equation for the rate of change of potential plus internal energy, it can be shown (see Chapter XVIII) that the diabatic terms containing ω_Q or $\frac{dQ}{dt}$ cancel out. Since we have measured only the adiabatic vertical motions, the term

$$\iiint \frac{dQ}{dt} \frac{dp}{g} dx dy \tag{16.1}$$

does not enter into the evaluation of our reduced equation.

On the other hand, the energy which drives the atmosphere first appears as diabatic heating.

From Murgatroyd and Singleton's (1961) article we obtained the following hemispheric heating rates.

Table16.1 Mean hemispheric stratospheric heating rates.

Summer	100 mb	$+.18^{\circ}$C/day
	50 mb	$+.20^{\circ}$C/day
Winter	100 mb	$+.15^{\circ}$C/day
	50 mb	$+.07^{\circ}$C/day

If we assume that $\frac{1}{c_p}\left\{\left[\overline{\frac{dQ}{dt}}\right]\right\}$ can be equated to a heating rate of $.1^{\circ}$C/day we obtain the average values of the integral (16.1) for the listed levels for the northern hemisphere.

Table 16.2 Values of $\iiint \frac{dQ}{dt} \frac{dp}{g} dx dy$ for a heating rate of .1°C/day.

100-50 mb	591 ergs cm^{-2} sec^{-1}
50-30 mb	236 "
30-10 mb	236 "

Under the above assumptions we may also estimate values of the mean term of the integral $\iiint \frac{\omega_Q \alpha}{g} dp\, dx\, dy$,

$$\omega_Q \alpha = \frac{\frac{R}{c_p} \frac{T}{p} \frac{dQ}{dt}}{\frac{R}{c_p} \frac{T}{p} - \frac{\partial T}{\partial p}} \approx \frac{dQ}{dt} \tag{16.2}$$

Comparison of values in Table 16.2 with the mean term of the integral

$$\iiint \frac{\omega_t \alpha}{g} dp\, dx\, dy \tag{16.3}$$

shows that both are the same order of magnitude (see Table 10.1).

Theoretical and observational evidence indicates that the diabatic and adiabatic vertical motions act against one another, so it seems that the values are of the right order.

Evaluation of the integral

$$\iiint \mathbb{V}_p \cdot \mathbb{F} \frac{dp}{g} dx\, dy \tag{16.4}$$

of equation (2.9) is very difficult. First it is necessary to discover the physical meaning of the term. We started with the horizontal equation of motion

$$\frac{d\mathbb{V}}{dt} + f\,\mathbb{K}\times\mathbb{V} + \nabla_p \psi + \mathbb{F} = 0 \tag{16.5}$$

The vector $\mathbb{F}$ is usually called the horizontal friction force. It would be more appropriate to consider it as the sum of all of the remaining terms which belong in the equation but which are not explicitly given. The integral (16.4) measures the dissipation of kinetic energy due to tidal effects as well as the dissipation due to hydromagnetic effects. Molecular frictional dissipation of kinetic energy would also be measured as a part of (16.4). However, it is usually considered that this latter energy reappears as heat, so we would expect this to be a transformation from kinetic energy to potential energy. This transformation term is included in the integral (16.1).

On the micro-scale we have dissipation of kinetic energy by what is called molecular friction. Meso-scale dissipation of the kinetic energy of the atmosphere can not be explained using only molecular friction. Generally, meso-scale dissipation is taken to include all rather small scale eddy effects (such as those found in the friction-layer) which deplete the energy of the atmosphere on this scale. Similarly we might expect to find that, on the macro-scale, the dissipation is again different than on the micro and meso-scales.

We would expect that, internally, there would be dissipation

of kinetic energy in the volume. On the other hand, if the stratosphere is dragged along by the troposphere, one might find an increase of kinetic energy due to the frictional stresses at the boundaries.

Values of the dissipation term for these layers could be given but they would be based on certain assumptions whose validity is questioned. The jump from meso-scale to macro-scale may give values as different as the jump from micro to meso-scale. Also the system is open, so frictional stresses could bring about an increase of kinetic energy (Starr, 1960).

We wish to summarize by saying that, for the dissipation more than the radiation, these integrals represent our present inability to express in a precise mathematical form (which could be easily used, and which would be amenable to calculation) certain physical properties concerning the energy of the atmosphere.

XVII Kinetic Energy of the Mean Zonal Wind

In this chapter we shall consider the equation for the time rate of change of the kinetic energy of the mean zonal wind. We shall evaluate some of the terms in this equation and draw some conclusions about the transfer of kinetic energy to the mean zonal motion.

The equation for zonal momentum in the λ, ϕ, p coordinate system may be written as:

$$\frac{\partial u}{\partial t} = -\frac{\partial(uv\cos^2\phi)}{a\cos^2\phi\,\partial\phi} - \frac{\partial u\omega}{\partial p} + fv - \frac{\partial(RT+u^*)}{a\cos\phi\,\partial\lambda} \qquad (17.1)$$

Taking the mean around a latitude circle, equation (17.1) becomes.

$$\frac{\partial[u]}{\partial t} = -\frac{\partial([uv]\cos^2\phi)}{a\cos^2\phi\,\partial\phi} - \frac{\partial[u\omega]}{\partial p} + f[v] \qquad (17.2)$$

If equation (17.2) is multiplied by $[u]$, the time mean is taken and integration over the northern hemisphere between two pressure levels is considered, then the rate of change of kinetic energy of the mean zonal motion for this layer may be written as:

$$\frac{2\pi a^2}{g(t-t_0)}\iiint_{t_0}^{t}\frac{\partial}{\partial t}\left(\frac{[u]^2}{2}\right)\cos\phi\;dt\,d\phi\,dp = \qquad (17.3)$$

$$+\frac{2\pi a}{g}\iint\overline{[u^*v^*]}\cos^2\phi\,\frac{\partial}{\partial\phi}\left(\frac{\overline{[u]}}{\cos\phi}\right)d\phi\,dp \qquad (17.4)$$

$$+\frac{2\pi a}{g}\iint \overline{[u]}\,\overline{[v]}\cos^2\phi\,\frac{\partial}{\partial\phi}\left(\frac{\overline{[u]}}{\cos\phi}\right)d\phi\,dp \qquad (17.5)$$

$$+\frac{2\pi a^2}{g}\iint 2\Omega\sin\phi\cos\phi\,\overline{[u]}\,\overline{[v]}\,d\phi\,dp \qquad (17.6)$$

$$+\frac{2\pi a^2}{g}\iint \overline{[u^*\omega^*]}\cos\phi\,\frac{\partial\overline{[u]}}{\partial p}\,d\phi\,dp \qquad (17.7)$$

$$+\frac{2\pi a^2}{g}\iint \overline{[u]}\,\overline{[\omega]}\cos\phi\,\frac{\partial\overline{[u]}}{\partial p}\,d\phi\,dp \qquad (17.8)$$

$$+\frac{2\pi a}{g}\int \overline{[uv]}\;\overline{[u]}\,dp \qquad (17.9)$$

$$+2\pi a^2\int\left(\overline{[u\omega]}_L\,\overline{[u]}_L-\overline{[u\omega]}_u\,\overline{[u]}_u\right)\cos\phi\,d\phi \qquad (17.10)$$

$$+\frac{2\pi a^2}{g}\iint\Bigg\{\;\overline{[u^*v^*]'\cos^2\phi\,\frac{\partial}{\partial\phi}\left(\frac{[u]'}{a\cos\phi}\right)}+\overline{[u]'[v]'}\cos^2\phi\,\frac{\partial}{\partial\phi}\left(\frac{\overline{[u]}}{a\cos\phi}\right)$$

$$+\overline{[u]}\cos^2\phi\;\overline{[v]'\frac{\partial}{\partial\phi}\left(\frac{[u]'}{a\cos\phi}\right)}+\overline{[v]}\cos^2\phi\;\overline{[u]'\frac{\partial}{\partial\phi}\left(\frac{[u]'}{a\cos\phi}\right)}$$

$$+\overline{[u]'[v]'\cos^2\phi\,\frac{\partial}{\partial\phi}\left(\frac{[u]'}{a\cos\phi}\right)}+2\Omega\sin\phi\cos\phi\,\overline{[u]'[v]'}$$

$$+\overline{[u^*v^*]'\cos\phi\,\frac{\partial[u]'}{\partial p}}+\overline{[u]'[\omega]'}\cos\phi\,\frac{\partial\overline{[u]}}{\partial p}$$

$$+\overline{[u]}\cos\phi\;\overline{[v]'\frac{\partial[u]'}{\partial p}}+\overline{[v]}\cos\phi\;\overline{[u]'\frac{\partial[u]'}{\partial p}}$$

$$+ \overline{[u]'[v]'\cos\phi \frac{\partial [u]'}{\partial p}} \Bigg\} d\phi\, dp \qquad (17.11)$$

$$+ \frac{2\pi a}{g} \int \overline{[uv]'[u]'}\, dp \qquad (17.12)$$

$$+ 2\pi a^2 \int \left(\overline{[u\omega]'_L [u]'_L} - \overline{[u\omega]'_u [u]'_u} \right) \cos\phi\, d\phi \qquad (17.13)$$

It should be noted, that in this chapter parenthesis, (   ), and braces, $\{\ \}$, are used only to separate parts of the equations and are not meridional and pressure means.

The integrals (17.4) through (17.10) were evaluated using our data. The integrals (17.11), (17.12) and (17.13) could not be evaluated from our data.

Evaluation of (17.3) was made by use of the geostrophic wind equation. Using daily maps of the height of the 100, 50 and 30 mb maps [prepared and published by the U.S. Weather Bureau (1960)], mean heights were obtained for each 5° of latitude for the maps at the beginning and end of both periods thus allowing the geostrophic $[u]$ to be computed for each 5° interval.

Since direct measurements of $\overline{[u^* v^*]}$ were not made, use of the identity

$$\overline{[u^* v^*]} \equiv [\bar{u}^* \bar{v}^*] + [\overline{u'v'}] - \overline{[u]'[v]'} \qquad (17.14)$$

and the empirical fact (Starr & White, 1952b) that $\overline{[u]'[v]'}$ is an order of magnitude smaller than $\overline{[u^* v^*]}$ suggested approximating (17.4) by the sum of the two integrals

$$\frac{2\pi a}{g}\iint [\overline{u}^*\overline{v}^*] \cos^2\phi \frac{\partial}{\partial\phi}\left(\frac{[\overline{u}]}{\cos\phi}\right) d\phi\, dp \tag{17.15}$$

and

$$\frac{2\pi a}{g}\iint [\overline{u'v'}] \cos^2\phi \frac{\partial}{\partial\phi}\left(\frac{[\overline{u}]}{\cos\phi}\right) d\phi\, dp \tag{17.16}$$

Since the order of integration is immaterial, the order of the space mean and latitude mean may be interchanged, e.g. $[\overline{u}] \equiv \overline{[u]}$

The $[\overline{V}]$ s were computed from the adiabatic vertical motions and from the actual winds. Integral (17.6) was evaluated using both sets of values. The noise level was very high. Integral (17.5) was evaluated using the adiabatic $[\overline{V}]$s with indications that the term is of minor importance. For this reason, the actual $[\overline{V}]$ s were not used to evaluate this integral. Also the actual $[\overline{V}]$ s contained too much noise.

The term

$$\frac{2\pi a^2}{g}\iint 2\Omega \sin\phi \cos\phi \; \overline{[u]'[v]'}\; d\phi\, dp \tag{17.7}$$

contained in integral (17.11) could not be evaluated from our data. A study by Starr and White (1952b) indicates that $\overline{[u]'[v]'}$ and $\overline{[u]}\overline{[v]}$ may cancel each other in the stratosphere. We assume that the remaining terms of (17.11) , (17.12) and (17.13) are small and may be disregarded.

The term (17.7) was treated in a manner similar to that used to

obtain values for (17.4).

$$\overline{[u^{*}\omega^{*}]} \equiv [\overline{u}^{*}\overline{\omega}^{*}] + [\overline{u'\omega'}] - \overline{[u]'[\omega]'}$$

No good comparison has been made of these four terms. We have computed only the first term on the RHS. The second term is being computed by other members of the Planetary Circulations Project. Preliminary indications (Loisel and Molla, 1961) are that the second term of the RHS is more important than the first, however we have values of only

$$\frac{2\pi a^2}{g}\iint[\overline{u}^{*}\overline{\omega}^{*}]\cos\phi\ \frac{\partial[\overline{u}]}{\partial p}\,d\phi\,dp \tag{17.18}$$

using the adiabatic vertical motions.

Table 17.1 gives the values of the integrals in units of 10^{17} ergs sec^{-1} mb^{-1}. The text number of the integral appears in the first column. The next three columns are for the 100, 50 and 30 mb levels respectively for the period July through September 1957. Integrals containing ω were computed for the two layers 100 to 50 mb and 50 to 30 mb, so the values were placed appropriately in the table. The last columns are for the period October through December 1957 in the same form as described above. Where both the 00Z and 12Z values were obtained the 12Z values are underneath the 00Z values. Adiabatic $[\overline{V}]$ values were not available at 100 mb. This is indicated by M in the table.

The largest values are associated with the integral (17.6) but this integral probably is balanced by (17.17) which was not

measured. Thus all we can say is that we can not disregard these two terms.

Apart from these "Coriolis" terms we find that the major contributions to the rate of change of the mean zonal kinetic energy are given by the terms (17.15) and (17.16) which approximate (17.4), the Reynold's stress term. These two terms are larger than any of the remaining terms.

Under the assumption that the unmeasured terms (except for (17.17)) are unimportant we conclude that the Reynold's stress terms and the Coriolis terms are the important terms in the equation for the rate of change of zonal kinetic energy in the stratosphere.

These conclusions seem to jibe with Phillips' (1956) finding for the upper level of his theoretical model of the general circulation.

Theoretically and observationally the short wave disturbances dampen with height in the stratosphere. On the other hand the transient eddies play a more important role than the standing zonal eddies in the stratosphere. This once again points up the fact that there is not a one-to-one correspondence between long and short waves and standing zonal and transient eddies.

Table 17.1 Northern hemisphere average values of the measured terms in the equation for the time rate of change of zonal kinetic energy. Units are 10^{17} ergs per second for a vertical layer of one millibar thickness.

Text Number	July - Sept. 100 mb	July - Sept. 50 mb	July - Sept. 30 mb	Oct. - Dec. 1957 100 mb	Oct. - Dec. 1957 50 mb	Oct. - Dec. 1957 30 mb	Remarks
(17.3)	+0	-2	-3	+7	+3	+3	Measured Change
(17.15)	+2 +2	-0 +0	-1 +1	+3 +2	+1 +0	-0 +1	Horizontal
(17.16)	+2 +3	-1 +0	-1 -1	+5 +4	+2 +1	+2 +2	Reynold's Stress Terms
(17.6)	M M	+7 +1	-1 +2	M M	-17 +6	-40 -34	Measured Coriolis Term Adiabatic $[\bar{V}]$
(17.6)	-28 +22	+13 -17	+2 -11	+45 +11	-11 -21	+3 -68	Measured Coriolis Term Actual $[\bar{V}]$
(17.9)	-0 -0	+0 +0	-0 +0	-0 -0	-0 +0	-0 -0	Boundary Term at Equator
(17.18)	+1 +1	+0 -0		+0 +0	+0 +0		Vertical Reynold's Stress Terms
(17.5)	M M	+1 +1	-0 -0	M M	-1 +0	-2 +1	Horizontal Mean Term
(17.8)	-0 -1	+0 +1		+0 +0	+1 +0		Vertical Mean Term

XVIII Conclusions

A. Rate of change of kinetic energy.

We rewrite equation (2.9) as

$$\frac{\partial}{\partial t}\iiint E \frac{dp}{g} dx dy = \iint E V \frac{dp}{g} dx + \iint \psi V \frac{dp}{g} dx - \iint \left(\frac{E\omega_c}{g}\right)_L dx dy$$
$$+ \iint \left(\frac{E\omega_c}{g}\right)_u dx dy - \iint \left(\frac{\psi\omega_c}{g}\right)_L dx dy + \iint \left(\frac{\psi\omega_c}{g}\right)_u dx dy - \iiint \omega_c \alpha \frac{dp}{g} dx dy$$
$$- \iint \left(\frac{E\omega_Q}{g}\right)_L dx dy + \iint \left(\frac{E\omega_Q}{g}\right)_u dx dy - \iint \left(\frac{\psi\omega_Q}{g}\right)_L dx dy + \iint \left(\frac{\psi\omega_Q}{g}\right)_u dx dy$$
$$- \iiint \omega_Q \alpha \frac{dp}{g} dx dy - \iiint \mathbb{V}_p \cdot \mathbb{F} \frac{dp}{g} dx dy \tag{18.1}$$

The first eight terms have been evaluated from the data. The next five depend on radiation and the last term is the friction term. Since the radiation and friction are not known well enough, these terms could not be evaluated. The radiation terms do not cancel out as they do in the equation for the time rate of change of potential plus internal energy.

The individual integrals of (18.1) are listed below along with our findings.

1. $\frac{\partial}{\partial t}\iiint E \frac{dp}{g} dx dy$ (Chapter IX) For all three layers there was a slight decrease fbom July through September and a net increase of kinetic energy from then to the end of December.

2. $\iint E V \frac{dp}{g} dx$ (Chapter XII) The

mean seasonal advection of kinetic energy across the equator in the lower stratosphere was small when compared with the mean seasonal time rate of change of kinetic energy in the stratosphere over the northern hemisphere. This integral is not important and vanishes when the whole sphere is considered.

3. $\iint \psi V \frac{dp}{g} dx$ (Chapter XIII) The eddy terms are not important, but the mean term may be.

4. $\iint \left(\frac{E\omega_c}{g}\right) dx\,dy$ and $\iint \left(\frac{E\omega_Q}{g}\right) dx\,dy$ (Chapter XI) The vertical advection of kinetic energy could account for the decrease of total kinetic energy from July through September as computed using the adiabatic vertical motion, but the October through December increase of kinetic energy of the stratosphere must have been due to some other process. Since ω_c is a function of the wind (through $\mathbb{V}_p \cdot \nabla_p T$) and ω_Q is not, one might expect the diabatic integral to be of less importance than the adiabatic one.

5. $\iint \frac{\psi\omega_c}{g} dx\,dy$ and $\iint \frac{\psi\omega_Q}{g} dx\,dy$ (Chapter XV There were difficulties in evaluating the adiabatic integral, but undoubtedly these terms are very important and cannot be disregraded.

6. $\iiint \mathbb{V}_p \cdot \mathbb{F} \frac{dp}{g} dx\,dy$ (Chapter XVI) At the present time this term cannot be evaluated, but it probably acts to decrease the kinetic energy of the stratosphere.

7. $\iiint \omega_c \alpha \frac{dp}{g} dx\,dy$ and $\iiint \omega_Q \alpha \frac{dp}{g} dx\,dy$ (Chapter X) Using adiabatic vertical motions, the transient and zonal standing eddies gave a conversion from kinetic to potential energy in both summer and winter in the lower stratosphere. Above about 40 mb the net conversion by transient and zonal eddies was in the opposite direction. The diabatic motions are important for the meridional eddy and mean terms, tending to cancel the

adiabatic contributions to these terms. Hence we do not know the true role of these terms in the rate of change of kinetic energy.

For the northern hemisphere the seasonal rate of change of total kinetic energy depended almost entirely on the terms in the last three numbered sections. This means that the large hemispheric increase of total kinetic energy observed in the early winter was not due to the vertical advection of kinetic energy.

B. <u>Rate of change of potential plus internal energy.</u>

From Chapter II we have

$$\frac{\partial}{\partial t}\iiint(\Phi+\mathcal{I})\frac{dp}{g}\,dx\,dy = \frac{p_L}{g}\iint\left(\frac{\partial\psi}{\partial t}\right)_L dx\,dy - \frac{p_u}{g}\iint\left(\frac{\partial\psi}{\partial t}\right)_u dx\,dy + \iiint c_p\frac{\partial T}{\partial t}\frac{dp}{g}\,dx\,dy \tag{18.2}$$

Using the first law of thermodynamics to obtain $c_p\frac{\partial T}{\partial t}$ we get

$$\frac{\partial}{\partial t}\iiint(\Phi+\mathcal{I})\frac{dp}{g}\,dx\,dy = \frac{p_L}{g}\iint\left(\frac{\partial\psi}{\partial t}\right)_L dx\,dy - \frac{p_u}{g}\iint\left(\frac{\partial\psi}{\partial t}\right)_u dx\,dy + \iint\frac{c_pTV}{g}\,dp\,dx$$
$$-\iint\left(\frac{c_pT\omega}{g}\right)_L dx\,dy + \iint\left(\frac{c_pT\omega}{g}\right)_u dx\,dy + \iiint\omega\alpha\frac{dp}{g}\,dx\,dy + \iiint\frac{dQ}{dt}\frac{dp}{g}\,dx\,dy \tag{18.3}$$

Since we have calculated the adiabatic vertical motion we have that

$$c_p\frac{\partial T}{\partial t} = -c_p\mathbb{V}_p\cdot\nabla_p T + c_p\omega_c\left(\alpha - \frac{\partial T}{\partial p}\right) \tag{18.4}$$

Substituting in (18.2) we find that all terms on the RHS can be evaluated with our data. $\frac{\partial}{\partial t}\iiint(\Phi+\mathcal{I})\frac{dp}{g}\,dx\,dy = \frac{p_L}{g}\iint\left(\frac{\partial\psi}{\partial t}\right)_L dx\,dy - \frac{p_u}{g}\iint\left(\frac{\partial\psi}{\partial t}\right)_u dx\,dy$

$$+\iint\left(\frac{c_pTV}{g}\right)dp\,dx - \iint\left(\frac{c_pT\omega_c}{g}\right)_L dx\,dy + \iint\left(\frac{c_pT\omega_c}{g}\right)_u dx\,dy + \iiint\omega_c\,\alpha\frac{dp}{g}\,dx\,dy \tag{18.5}$$

Notice that no diabatic terms (i.e. containing ω_Q or $\frac{dQ}{dt}$) appear in (18.5). The $[\overline{V}]$ and $[\overline{\omega_t}]$ values will not satisfy the continuity equation unless $[\overline{\omega}] \doteq [\overline{\omega_t}]$ for all latitudes which means that the $[\overline{\omega_Q}]$ values must be at least an order of magnitude smaller than the $[\overline{\omega_t}]$ values. Chapter VIII indicates that this is not true.

Since the radiation terms cancel one another, we do not expect to learn much about radiation in the stratosphere using the equation for the rate of change of potential plus internal energy and adiabatic vertical motions.

Listed below are the various terms of (18.5) along with conclusions from previous chapters.

1. $\frac{\partial}{\partial t}\iiint(P+I)\frac{dp}{g}\,dx\,dy$ (Chapters IX) There is a net decrease of potential plus internal energy at all levels in the northern hemisphere stratosphere during both periods. This is due to a net lowering of the pressure surfaces and decrease of the mean temperature.

2. $\iint\left(\frac{c_p T V}{g}\right)dp\,dx$ (Chapter XIV) This boundary term is evaluated at the equator. The eddy terms are not important contributors to the time rate of change but the mean term is. Unfortunately, evaluation of the mean term was not considered reliable with our data. This integral does not appear if we integrate over the entire sphere.

3. $\iint\left(\frac{c_p T \omega_t}{g}\right)dx\,dy$ (Chapter X) The meridional eddy term and mean term are important, but the other eddy terms do not contribute significantly.

4. $\frac{f}{g}\iint \frac{\partial \psi}{\partial t}\,dx\,dy$ (Chapter IX) This is a very important term. The major change of potential energy was due to the change of the mean height of the lower stratosphere rather then redistribtuion of mass within the lower stratosphere.

5. $\iiint \omega_c \alpha \frac{dp}{g}\,dx\,dy$ (Chapter X) Except for the layer from 30 to 10 mb, the eddy terms did not contribute significantly. On the other hand, the mean terms were very important and cannot be disregarded in our equation for the rate of change of potential plus internal energy.

The remaining three terms are all functions of $\frac{dQ}{dt}$ and are all zero under the adiabatic assumption. However, as indicated above, this does not mean that one is measuring the rate of change of potential plus internal energy in an adiabatic atmosphere. In an adiabatic atmosphere, $\frac{dQ}{dt} = 0$, and the kinetic energy goes to zero with time. This does not happen in the real atmosphere from which we are taking measurements.

Using adiabatic vertical motions, the rate of change of potential plus internal energy appears to be a small difference among large terms. These large terms all stem from the integral $\iiint c_p \frac{\partial T}{\partial t} \frac{dp}{g}\,dx\,dy$. Thus this method of separating $c_p \frac{\partial T}{\partial t}$, as in equation (18.4), does not seem to be a good way of evaluating this term.

C. <u>Rate of change of the zonal kinetic energy.</u>

From Chapter XVII we see that the kinetic energy of the 1957 stratospheric zonal winds was derived primarily from two sources, the

Reynolds's stress terms and the Coriolis transformation terms. At 100 mb for both periods the eddy stress terms (17.15) and (17.16) provided more than enough kinetic energy. Not all the terms were evaluated, but the indications were that the meridional circulation acted to deplete the kinetic energy of the 100 mb zonal winds through the Coriolis terms, (17.6) and (17.17). During the period July through September 1957 at 50 and 30 mb, depletion of zonal kinetic energy by the Coriolis terms was necessary to give the observed change of zonal kinetic energy. The Coriolis term must have operated in the opposite direction to increase the zonal kinetic energy at 50 and 30 mb for the October through December period to give the observed changes. These conclusions, derived from the actual data, were indicated in a two level numerical model investigated by Phillips (1956). This is related to the fact that the zonal flow changes from easterly to westerly. We have assumed that the vertical eddy term

$$\frac{2\pi a^2}{g} \iint \overline{[\omega^* u^*]} \frac{\partial [u]}{\partial p} \cos\phi \, d\phi \, dp$$

actually can be disregarded. Preliminary evidence indicates that it can be disregarded for our purposes.

Thus, for the northern hemisphere lower stratosphere (below roughly 75 mb), we found a conversion of eddy kinetic energy into zonal kinetic energy and zonal kinetic into potential plus internal energy. During the July through September period the total kinetic energy of this layer remained almost constant. Therefore the large amount of zonal kinetic energy supplied by the eddy kinetic energy through the Reynolds stress must have been dissipated by the meridional circulation through

the Coriolis terms. This means that the eddy kinetic energy became potential plus internal energy while the zonal kinetic energy acted as an almost unchanged intermediate form of energy.

On the other hand, in the lower stratosphere from October to December 1957, the zonal kinetic energy increased rapidly at the expense of the eddy kinetic energy but with a small conversion to potential plus internal energy.

Above this level, in the upper stratosphere (75 mb to 30 mb), the Reynolds' stress decreased the zonal kinetic energy during the summer and increased it during the early winter. To obtain the observed change in zonal kinetic energy the meridional circulation must have brought about a decrease in the zonal kinetic energy (when acting through the Coriolis term) during the summer and an increase during the early winter.

We conclude that the eddy kinetic energy is one of the major sources of zonal kinetic energy, but another source must exist. The evidence points to potential energy as the other source. Thus it appears to us that the meridional circulation plays a more important role in the energy balance of the stratosphere than it does in the energy balance of the troposphere where the eddy kinetic energy is more nearly the sole source of zonal kinetic energy.

This has been a preliminary study of the rate of change of zonal kinetic energy in the stratosphere and more work is needed. Mr. S. Teweles is studying this problem using the daily maps. His approach will simplify the equation and allow more terms to be evaluated.

D. The importance of eddy transports in the stratosphere.

Eddy transport of various quantities in the stratosphere can not be disregarded. Historically one finds a similarity between the sequences of concepts concerning the transport processes in the stratosphere and in the troposphere. Models using only meridional circulations still are being discussed in the literature, but theoretical and observational investigations indicate that eddy transports are important and that the actual atmosphere is a combination of the symmetric and eddy regimes.

In the troposphere the measured eddy transports are capable of providing almost all the necessary transports, and the transports by the mean meridional circulation pattern seem to be of secondary importance. This paper provides empirical evidence that both the mean meridional circulation pattern and the horizontal eddies are necessary for a description of the energy balance of the stratosphere.

XIX Discussion

Since we are working under the theme that the atmosphere is its own best model what one needs is a good four-dimensional picture of the atmosphere. This study has shown that we need more and better data over the whole sphere at all levels. We have had to exclude all of the southern hemisphere as well as the atmosphere above the 10 mb level from this study because of the sparcity of stratospheric data.

Standardization of upper air sounding instruments would be a great boon to this type of study. The derived heights of the pressure levels in the stratosphere are so erratic that it is necessary to employ the mean wind pattern to deduce the mean height pattern at high levels. Unfortunately, political pressures have played as significant a role in the selection of sounding instruments used by various countries as have scientific and technological considerations.

With the recent advances in analysis by means of high-speed computers, it will be necessary in the future to conduct studies such as this one completely automatically. In other words, the analysis of individual maps should be done by the computer, the grid point values should be taken by the computer and only the final values should be printed. At the discretion of the investigator, intermediate steps, to be used for other purposes, could be made available.

Using the 3 month average of the adiabatic vertical motion, we obtained the adiabatic meridional circulation for the stratosphere. The agreement between the 00Z means and the 12Z means indicates that for two three month periods we successfully obtain the reliable adiabatic

meridional motions. For the actual mean meridional motions, as shown by the $[\overline{V}]$ s, we found that the signal was not separated from the noise for a 3 month period, but, by using six months of data for both 00Z and 12Z at 50 mb over the northern hemisphere, we obtained a fairly smooth curve which agreed with Jensen's circulation for January 1958. The 6 month mean $[\overline{V}]$ s at 100 mb contained too much noise due to the proximity of the high winds just below the tropopause. The 6 month $[\overline{V}]$ s at 30 mb could not be separated from the noise due to the lack of a sufficient number of soundings reaching this level during the 6 month period. Using 12 or 18 months of data, we hope to obtain the actual $[\overline{V}]$ s at 100 mb and 30 mb.

At the 50 mb level the adiabatic $[\overline{V}]$ s for the 6 month period were subtracted from the real $[\overline{V}]$ s leaving the diabatic $[\overline{V}]$ s. These diabatic motions have smaller magnitudes but, in general, the same direction as the horizontal meridional motions calculated by Murgatroyd and Singleton (1961) which are believed to be weighted very heavily towards mean diabatic (vs adiabatic) motions, We have shown that Murgatroyd and Singleton's disregard of the eddy terms is not permissable when trying to compute the meridional circulation the way they did.

Generally, in the lower stratosphere, diabatic and adiabatic motions opposed each other, and we found that the 6 month mean $[\overline{V}]_c$ s and $[\overline{V}]_Q$ s were of comparable magnitudes. This points up a very important feature of the atmosphere, Namely, the diabatic motions can not be disregarded when dealing with long period mean motion in the stratosphere.

A study of the term $[\overline{\omega_Q}]$ indicates that the transient eddies in the radiation field were as important as the mean terms in determining

the mean vertical diabatic motions. This would seem to indicate that any long range forecasting technique which takes radiation into account would have to include the temporial variations of the radiation throughout the atmosphere as well as the mean radiation in the atmosphere.

Our evaluation of the equation of the mean time rate of change of the zonal kinetic energy indicates that the zonal kinetic energy changed primarily due to the Reynolds' stress terms and the meridional motions acting through the Coriolis terms. In other words the primary sources and sinks of zonal kinetic energy in the stratosphere were the eddy kinetic energy and the gradient of the meridional potential energy.

Investigation of the mean time rates of change of the total kinetic energy and mean time rate of change of the total potential plus internal energy for the two periods indicates that the upward flux of kinetic energy during the summer was sufficient to account for the change of kinetic energy in the 100 to 50 mb layer during the summer, but the conversion of the potential plus internal energy to kinetic energy in situ may have been as important.

During the winter the vertical flux of kinetic energy was not large enough to account for the observed changes in the kinetic energy of the stratosphere.

In the lower stratosphere the transient eddies and standing zonal eddies converted kinetic energy to potential plus internal energy. In the upper stratosphere we found a conversion from potential plus internal energy to kinetic energy just as previously found in the troposphere by other investigators. We must remember though that these

conclusions were for the stratosphere in 1957 over the whole northern hemisphere and that over certain regions at certain times this mean picture did not apply.

Studies of ozone and radioactive debris in the stratosphere indicates that, in the mean, these tracers move approximately parallel to the mean isentropic surfaces. Newell (1961) has shown that these tracers may move from equatorial region to about 50°N and 50°S by "shuffling" or eddy motion. This is consistent with White's (1954) findings, which we have reconfirmed, that there is a countergradient heat flux in the lower stratosphere. Air which has been heated by adiabatic sinking will also be richer in ozone and radioactive debris since the source regions are above the lower stratosphere. The ωT correlations show that the transient and zonal eddies transport heat downward so we might also expect that the tracers would be brought downwards. Correlations of V and ω by Loisel and Molla (1961) show that for January and April 1958 sinking air generally moves northward in the layer from 100 to 50 mb.

Studies of tropospheric-stratospheric interchanges have been successful using the concepts of constant potential vorticity and constant potential temperature. Thus the above findings do not seem to be out of line. On the other hand we have shown indications that the diabatic heating is important when considering the meridional circulation. This probably accounts for some of the discrepancy between the axis of the spread of the tracers and the isentropes, but gravitational settling must be considered too.

With the source of water vapor being below the lower stra-

tosphere, we would expect to find an equatorward flux of water vapor in the lower stratosphere. So far the only piece of information which would help us in this respect is the observed negative correlation between ozone and water vapor in the stratosphere.

In retrospect, the major role played by radiation in the mean circulation of stratosphere (as shown by the diabatic $[\overline{V}]$ s) should not have been as startling to us as it was because we know that the atmosphere is driven by the solar radiation and if it were not for this radiation there would be no circulating atmosphere.

Investigations in the field of atmospheric radiation indicated that the lower stratosphere is probably that layer of the atmosphere where the adiabatic assumption is most closely satisfied. At lower levels, particularly near the ground, the transfer of solar radiation to the atmosphere is relatively large. In the ozone layer again there is relatively strong absorption of solar radiation, and the presence of clouds in the troposphere gives an excess of terrestrial radiation emitted from this lower region in the atmosphere.

Under certain assumptions concerning the relative magnitudes of the temporial variations, the time means and the space means of the time means of $\frac{dQ}{dt}$, we have shown that the temporial variations of $\frac{dQ}{dt}$ are very important in determining the mean meridional diabatic motions in the lower stratosphere. Using these assumptions we found that the instantaneous adiabatic motions were generally much larger than the instantaneous diabatic motion. Assuming that the time covariances of the diabatic vertical motions also were smaller than those with the adiabatic vertical motions, we used the adiabatic vertical motions for

calculating transient eddy terms containing the vertical motion.

A similar analysis of the 3 month means indicated that the major features of the maps of the mean adiabatic vertical motions represented the major features of the maps of the mean actual vertical motions. Assuming that the covariances (around the latitude circles) of the mean diabatic vertical motions were smaller than those with mean adiabatic vertical motions, we used the 3 month mean adiabatic vertical motion fields for the actual mean vertical motion fields when calculating the standing zonal eddy terms.

On the other hand, the latitudinal means of the time means of the diabatic and adiabatic motions were of the same magnitude. Hence we did not feel justified in disregarding the diabatic vertical motions when calculating the meridional standing eddy terms or the mean terms over the northern hemisphere.

The importance of the transient eddies in determining the mean meridional diabatic motions of the stratosphere is very disturbing. If the eddies in $\frac{dQ}{dt}$ are as important for the mean motion of the troposphere as they are in the stratosphere, then we would expect that long range forecasting must include an estimate of the temporal and spacial variations of $\frac{dQ}{dt}$ if any measure of radiation is to be considered as a parameter in making the forecast.

In "Analysis of Satellite Infrared Radiation Measurements on a Synoptic Scale," Weinstein and Suomi (1961) found indications of a relationship between the infrared radiation centers and the corresponding surface low and high pressure centers for the area $50^{\circ}N$ to $40^{\circ}N$ in the western hemisphere when in the shadow zone portion of the

earth. Thus we would like to suggest using the pressure pattern as feedback to give the radiation pattern in numerical forecasting.

XX Suggestions for Further Research.

Some of these suggestions stem directly from this work while others concern basic meteorological questions which are related to topics considered in this paper.

In order to check for consistency of the data and analysis, maps and various averages of the quantity $\overline{C^2}$ should be compared with maps and averages of the mathematically equivalent quantity,

$$\overline{u}^2 + \overline{v}^2 + (S(u))^2 + (S(v))^2$$

This was not done here because the maps and averages of $S(u)$ and $S(v)$ were not available.

As a by-product to this paper, we have partially processed most of the quantities necessary for studies of the sensible heat balance and angular momentum balance of the stratosphere. Other members of the Planetary Circulations Project already have begun a study of the angular momentum balance using our data.

Ozone is one of the important tracers used in the study of the circulation of the stratosphere. Newell (1961) has used some of our data to investigate the horizontal motion of ozone in the stratosphere. We have written computer programs to study the vertical eddy flux of ozone employing the raw data tapes used in this thesis. The vertical movements of ozone should be studied to help confirm or refute various theories on ozone movements.

A more detailed study of the adiabatic vertical motions in the equatorial regions should be undertaken. Our method of computing the adiabatic vertical motion requires the use of the thermal wind which is

latitude dependent and makes $V_p \cdot \nabla_p T$ approach zero at the equator. This problem may be circumvented by drawing daily maps to obtain $V_p \cdot \nabla_p T$ in equatorial regions from direct measurements. However the problem of available data then becomes a stumbling block.

In past years we have witnessed the heroic deeds of men who have journeyed to the poles of the earth to obtain meteorological and other scientific data to further man's search for knowledge. We would like to suggest a much less heroic journey which would be equally important scientifically. The establishment of first class rawinsonde stations on islands in the eastern equatorial Pacific would be a great help to empirical studies such as this one and for studies of the equatorial region per se.

The use of spherical coverage instead of hemispheric coverage would eliminate the equatorial boundary terms in the equations studied. Also the mean terms in $\{[\overline{\omega}]\}$ become identically zero.

The importance of the horizontal eddies in the circulation of the atmosphere has been demonstrated by Starr and others. However the importance of the pressure eddies, defined in Chapter III, has never been established because of the small number of pressure levels used in previous studies. Future studies of the atmosphere at, say, every 50 mb from 1000 mb to 50 mb should be used to investigate these pressure eddies.

Calculations of the $[\overline{V}]$ s over 12 and 18 month periods should be carried out. For these periods one should probably be able to separate the signal from the noise at the 100 mb and 30 mb levels as well as at the 50 mb level to obtain the actual $[\overline{V}]$ values. Using these and the

adiabatically derived $[\bar{V}]$ values one could again deduce the mean horizontal meridional diabatic motions to learn more about the role of radiation in the mean circulation of the stratosphere.

The study of the mean time rate of change of the zonal kinetic energy in the stratosphere as carried out in this paper did not consider all of the terms of the equation since we took the time mean before the space mean. Evaluation of the terms taking the space means before the time means will be possible using the U.S. Weather Bureau stratospheric maps. It is understood that Mr. S. Teweles of the Weather Bureau is planning to evaluate the terms of this equation employing the geostrophic assumption to obtain the winds.

The importance of $\frac{dQ}{dt}$ and especially its time variations has been pointed out in previous chapters. Recent work by Weinstein and Suomi (1961) indicates that the infrared radiation detected by satellites is highly correlated with the cloud cover in the region from $40^{\circ}N$ to $50^{\circ}N$ in the western hemisphere at night. The cloud coverage is related to the surface pressure pattern, hence it is suggested that the predicted pressure pattern might be used, with certain statistically derived weighting functions, to represent the pattern of $\frac{dQ}{dt}$ for inclusion in dynamical prediction models.

The need of better methods of predicting cloud patterns is absolutely necessary, since the above suggestion would be only a stopgap method of employing radiation in prediction models.

When this study was begun, we realized that there were systematic mean inter-diurnal variations in the quantities we were dealing with in the stratosphere. Some of these variations are primarly instrumental, such as those in temperature, while others are real. We found evidence of such a real variation in the winds at both of two stations we investigated in detail. Programs have been written by the author for a hemispheric study of these wind variations using the basic data tapes employed in gathering this paper material. This work will be continued by other members of the Planetary Circulations Project.

Because of these inter-diurnal variations, the 00Z and 12Z observations were treated separately to facilitate further studies of the inter-diurnal variations of the quantities.

The following 12 months of the IGY will be studied in the same manner as we have done here to reconfirm or change what we have found and to show further the seasonal variation in the energy balance.

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MOMENTUM BALANCE OF THE STRATOSPHERE DURING THE IGY

by

Robert Earl Dickinson

ABSTRACT

A differential form of the equation of angular momentum balance is given and data from one year of actual wind observations in the stratosphere are used to discuss the terms involved. Horizontal eddy momentum convergences are largest at 100 mb near the maximum of the tropospheric jet, and at 50 and 30 mb in the vicinity of the polar night jet. The momentum divergences south of 30°N at 50 mb and 30 mb are sufficiently small to show that the seasonal changes of momentum cannot be due only to these eddy fluxes. No process could be found to give a stratospheric momentum balance if a northward meridional circulation in middle latitudes were assumed to exist.

I. Introduction

The MIT Planetary Circulations Project has processed hundreds of thousands of individual wind and temperature statistics for the stratosphere during the IGY. Several hundred hemispheric maps have been drawn to reduce station data to usable form. Yearly means of quantities were obtained by adding the results of maps drawn separately for 00Z and 12Z and for four different three-month time periods so that they can be considered much more significant statistically than earlier results obtained by drawing a single map for yearly data as

was done by Buch (1954), primarily for the troposphere.

This improved data has suggested the use of a differential representation to describe the manner in which the atmosphere fulfills the conservation of angular momentum at each point in the two-dimensional plane with northward distance and pressure the independent variables and various mean terms and correlations as dependent variables. The availability of increasingly better data should allow similar study of other questions. The suggested representation is used to provide a detailed description of the momentum balance of the northern hemisphere stratosphere for one year of actual wind data. As an application of tho momentum data, it is demonstrated that it is difficult to reconcile with any known dynamical process a poleward mean meridional motion in middle latitudes between 100 mb and 30 mb. The directly measured southward motion (Oort 1962) need not be used to reach this conclusion.

II. Previous Data Studies

In recent years an understanding of the operation of the general circulation of the atmosphere has been obtained by reduction of large amounts of meteorological data to a few suitably chosen statistics. The basic physical laws that must operate in any mechanical and thermodynamical system were used in order to give a well-defined significance to the results. This work has been aided and encouraged by analytical work and numerical experiments which can examine more exactly certain simplified models of the general circulation based on the actual atmospheric data. Likewise rotating tank experiments have provided information on the universality of the laws that hold in the earth's atmosphere for certain rates of rotation and heating.

The general circulation thus is being studied in order to describe the statistical manner in which instantaneous and mean local disturbances plus mean motions contribute to the fulfillment of basic physical laws. By means of semidaily radiosondes or other forms of observations, one defines a time dependent meteorological field in three dimensional space, consisting of the velocity field and the scalar fields of temperature, pressure, humidity or of other quantities of interest. Details too small in time and space to be distinguished by the observational network are ignored except when their effects on the large-scale phenomena need be taken into account through some form of parametric representation. The general circulation studies of the last decade have transformed the dependent variables of the three-dimensional meteorological field (which has a time scale from hours to days) to

dependent variables of a two-dimensional climatological field (which has a time scale of months to years.) The methods used are time and space means, the space mean being the average around a latitude circle. As the equations which govern the meteorological field are quadratic in form, the average correlations of these dependent variables become important in any set of equations describing the evolution of the climatological field. Information on the magnitude of these quantities, designated as the eddy terms, has been obtained in recent years. Present evidence suggest that time averages will not be strictly the same for different years, but it appears that the annual changes are not as large as the seasonal ones.

Previous momentum studies by Starr and White (1952), Buch (1954), and Mintz (1955) demonstrated that the frictional gain of zonal momentum from the ground in low latitude easterlies is transported to middle latitudes by eddy fluxes, whence it is dissipated by the mean surface westerlies. These conclusions were reached by integration over the pressure coordinate. By this means, lack of smoothness in the data at individual pressure levels became insignificant, and it was unnecessary to discuss the fashion in which momentum was transported vertically. This integral approach to general circulation studies has been very valuable in studying the atmosphere when more detailed information could not be obtained.

Geostrophic winds used by Mintz (1955) eliminate small-scale

noise so prevalent in direct wind observations; therefore they are the best possible means of studying atmospheric processes over short time periods. When averaged over a sufficiently long time period, perhaps in the order of a month, direct wind measurements should become as meaningful as geostrophic winds. Geostrophic momentum fluxes leave systematic errors discussed by Lorenz (1954). They may also suffer from systematic smoothing of the smaller scale disturbances that carry momentum, especially in regions of sparse data coverage. The present methods of using direct winds for general circulation studies also suffer several limitations. Random errors introduced into the analysis of a hemispherical map may be significantly large. Although in the last few years, radiosonde stations have become sufficiently extensive to give fairly complete hemispherical data coverage for the northern hemisphere, nevertheless, there are a number of regions of sufficiently sparse data coverage such that subjective analysis can introduce systematic errors affecting the zonal means of quantities with large zonal variations. Discretization errors of the numerical methods used place an ultimate limit on the accuracy obtainable.

In addition to the studies mentioned above, and extensions of them, there is found in the literature much other work that has been done on the global momentum balance, but because of inadequate data, it should be regarded as inconclusive. These studies have frequently been based on measurements of meridional circulations lacking data around a complete latitude belt. Meridional circulations in general can be expected to be of much smaller magnitude than the north-south motion at

any one grid point, the latter of which is predominately of wave numbers other than 0, so any lack of complete zonal data coverage may give meaningless results with systematic errors far exceeding the quantity being measured. In general the zonal variation of most climatological variables is sufficiently large that results derived from incomplete data coverage can be greatly in error.

III. Notation

(x,y,p) = coordinates in eastward, northward directions and pressure coordinate

t = time

$\vec{c}$ = (u,v,ω) = time rates of change of x,y,p following the motion

ϕ = latitude

λ = longitude

f = Coriolis parameter = $2\,\Omega\,\sin\phi$

θ = potential temperature

Q = nonadiabatic rate of heating per unit mass

a = radius of the earth

Ω = rate of rotation of the earth

$\overline{(\quad)} = \frac{1}{t_2-t_1}\int_{t_1}^{t_2}(\quad)\,dt$ = time average

$(\quad)' = (\quad) - \overline{(\quad)}$ = deviation from time average

$[(\quad)] = \frac{1}{2\pi}\int_0^{2\pi}(\quad)\,d\lambda$ = zonal average of one map

$(\quad)^* = (\quad) - [(\quad)]$ = deviation from a zonal average

$\widetilde{[(\quad)]} = \frac{1}{N}\sum_{i=1}^{N}[(\quad)]_i$ = mean zonal average of several maps

$[(\quad)]'' = [(\quad)] - \widetilde{[(\quad)]}$ = deviation of zonal average of one map from the mean zonal average

IV. Equations

The conservation of absolute zonal angular momentum in an arbitrary volume V enclosed by a surface S may be written

$$(1) \qquad \iiint_V \frac{\partial M}{\partial t}\, dV = -\iint_S M \vec{c}\cdot\vec{n}\, dS + \iiint F_x\, a\cos\phi\, dV$$

where $\quad M = (u + a\Omega\cos\phi)\, a\cos\phi$

is the absolute angular momentum about the earth's axis.

F_x = body forces acting about the earth's axis including pressure gradients and friction.

We use the divergence theorem, equation of continuity, time and zonal averages defined in (III), and assume our volume is above the highest mountains to get:

$$(2) \qquad \iint_{\phi\, P} a\cos\phi\, \Bigg(\overline{\frac{\partial[u]}{\partial t}} + \frac{1}{a\cos^2\phi}\frac{\partial \cos^2\phi}{\partial\phi}\left([\overline{u'v'}] + [\bar{u}^*\bar{v}^*]\right) + \frac{\partial}{\partial p}\left([\overline{u'\omega'}] + [\bar{u}^*\bar{\omega}^*]\right) \; [\bar{v}]\frac{\partial[\bar{u}]\cos\phi}{a\cos\phi\,\partial\phi}$$

$$[\bar{\omega}]\frac{\partial[\bar{u}]}{\partial p} - f[\bar{v}] - (Friction)_x \Bigg)\, dP\, a\, d\phi = 0$$

By friction we mean $\dfrac{1}{a\cos^2\phi}\dfrac{\partial}{\partial\phi}\cos^2\phi\,[\overline{uv}] + \dfrac{\partial}{\partial p}[\overline{u\omega}]$

on a scale smaller than the synoptic scale, as molecular friction is almost completely negliglible in the free atmosphere.

We note the term

$$(3) \qquad \int_{\phi_1}^{\phi_2}\int_{P_1}^{P_2} -f[\bar{v}]\, a\cos\phi\, dp\, a\, d\phi = \left[\int_S (\Omega a^2\cos^2\phi)\; \vec{c}\cdot\vec{n}\; dS\right]$$

so that the flux of the earth's absolute angular momentum, sometimes called Ω angular momentum, can also be interpreted as a torque produced by the Coriolis force acting on the volume. This first form has greater ease of computation, as one only needs to know the $[\bar{v}]$ while the other notation requires $[\bar{v}]$ and $[\bar{\omega}]$. This term can not produce any net horizontal momentum transport, as $[\bar{v}]$ integrated over the pressure coordinate is zero for a long time mean. However, it may be a very important mechanism for vertical momentum transport. For instance, a mean southward motion in middle latitudes above 850 mb could account for all the momentum convergence of 1950 reported by Buch (1954) or Starr and White (1954), if $[\bar{v}]$ averaged with respect to pressure was approximately 5 cm/sec.

Since the integral vanishes for an arbitrary volume, the integrand must vanish for each point in the volume. Thus one gets

$$(4) \qquad a\cos\phi\left(\frac{\partial\overline{[u]}}{\partial t} + \frac{\partial}{a\cos^2\phi\,\partial\phi}\left(\cos^2\phi\,[\overline{u'v'}] \quad [\bar{u}^*\bar{v}^*]\right) + \frac{\partial}{\partial p}\left([\overline{u'\omega'}] + [\bar{u}^*\bar{\omega}^*]\right)\right.$$

$$\left. + [\bar{v}]\,\frac{\partial[\bar{u}]\cos\phi}{a\cos\phi\,\partial\phi} + [\bar{\omega}]\,\frac{\partial[\bar{u}]}{\partial p} - f[\bar{v}] - \text{Friction}\right) = 0$$

This equation expressing the conservation of angular momentum at each point in the (y,p) space is of course but a form of the equation of east-west motion in spherical coordinates. The Coriolis torque term is entirely equivalent to the net convergence of angular momentum at any point in the (y,p) space.

The purpose of this paper is to discuss the measurements of the various terms in the above zonal momentum equations. The factor a $\cos\phi$ is dropped so that the units will be ones of simple acceleration: cm/sec^2. Thus a clear picture can be preserved of the effects of the various terms on the atmospheric dynamics.

V. The Data

The data used in this study have been and will be discussed in detail in other publications, for instance: Barnes (1962) and Murakami (1962). Thus it suffices to give a brief summary of their source. One year's actual stratospheric winds (July 1957 through June 1958) have been processed by the MIT General Circulations Project for a northern hemisphere network of stations. For the first six months the stations used are those described in Murakami (1962) 212 in number. For a second six months about twenty new stations were found to aid the analysis in equatorial regions, and three stations were added in the arctic on the Russian side of the north pole. Maps were analysed from these station data. Mr. A. Oort analysed maps of $\bar{v}$ and $\bar{\omega}$ for 00Z and 12Z for three monthly periods of the second six months at 100 mb, 50 mb and 30 mb (12 maps for each quantity). They are discussed by Oort (1962). The author analyzed maps of $\bar{u}$, $\overline{u'v'}$, and $\overline{u'\omega'}$ for the same period. The maps used for the first six months were the $\bar{u}, \bar{v}$, and $\overline{u'v'}$ presented by Murakami (1962) except that it was necessary to reanalyze the area over Russia because of a systematic error found in some of the material. The quantities $\bar{u}$, $\bar{u}^*\bar{v}^*$, $\bar{u}^*\bar{\omega}^*$, $\bar{u}'\bar{\omega}'$ were obtained by finite difference methods using a grid of 5 degrees of longitude by 5 degrees of latitude. This provided zonal averages of each quantity of interest for 00Z and 12Z each three-monthly time period. The results were added to get seasonal and yearly means. The term $\widetilde{[\bar{u}]''[\bar{v}]''}$ was investigated and found insignificant. Thus we may assume that $\widetilde{[\bar{u}][\bar{v}]} = \widetilde{[\bar{u}]}\,\widetilde{[\bar{v}]}$. The random errors in measuring $[\bar{v}]$ on one map were sufficiently great that no

significance can be attached to such correlation as the above, although a small actual correlation can be expected from seasonal changes in the $[\bar{u}]$ and $[\bar{v}]$.

In general it can be said that the data available for this study were much more extensive that that used for any previous study of stratospheric dynamics.

VI. Momentum Divergence by Horizontal Eddies

The term $\frac{1}{a\cos^2\phi}\frac{\partial}{\partial\phi}\cos^2\phi\,[\overline{u'v'}+\overline{u}^{*}\overline{v}^{*}]$ was evaluated by finite difference methods, using $2\Delta = 10^{\circ}$ latitude (see Table I; figure 1, figure 2, and figure 3). No attempt was made at smoothing except in drawing graphs so that it is indicative of the quality of the data that rather smooth results were obtained even after the process of differentiation. A similar computation with previously reported horizontal fluxes at 100 mb, for instance Buch (1954), would give wildly oscillating values.

At 100 mb there is a maximum convergence in the vicinity of the climatological troposphere jet stream near 35° North (see figure 4). During the summer the maximum convergence is at 35° N, 5° south of the jet stream, while in winter it is also at 35° N, 5° north of the jet stream. The close correlation between the mean zonal wind and momentum convergence can be explained as resulting from barotropic stability of the mean zonal flow in a statistical sense. The eddy terms act to maintain the mean zonal flow. Both the momentum convergence and the mean zonal flow is approximately twice as large in winter as in summer, corresponding somewhat to earlier results of Starr and White (1951) for the troposphere.

At 50 mb and 30 mb the momentum convergence during the winter is maximum in the region of the polar night jet (see table 1). It is also to be noted that in the equatorial stratosphere the eddy momentum divergence at 50 mb and 30 mb is much smaller than that at 100 mb. The eddies are also not as strong as shown by smaller $[\overline{\sigma(u)}]$ and $[\overline{\sigma(v)}]$ at these levels presented by Murakami (1962). It will be shown in Chapter IX

that the eddy divergences in the equatorial stratosphere are a second order effect at 50 mb and 30 mb in comparison to seasonal changes of the zonal momentum $\overline{\frac{\partial [u]}{\partial t}}$.

The momentum divergence in low latitudes and convergence in middle latitudes obtained at 100 mb are similar to that of the upper troposphere as indicated by momentum fluxes computed in earlier studies, for instance Starr and White (1954). The data at 50 mb and 30 mb indicate significant differences from tropospheric behavior.

The divergences of standing eddies, $\bar{u}^*\bar{v}^*$, and transient eddies $\bar{u}'\bar{v}'$ are not presented separately. The standing eddies while somewhat smaller, are of the same magnitude as the transient eddies and act in the same sense.

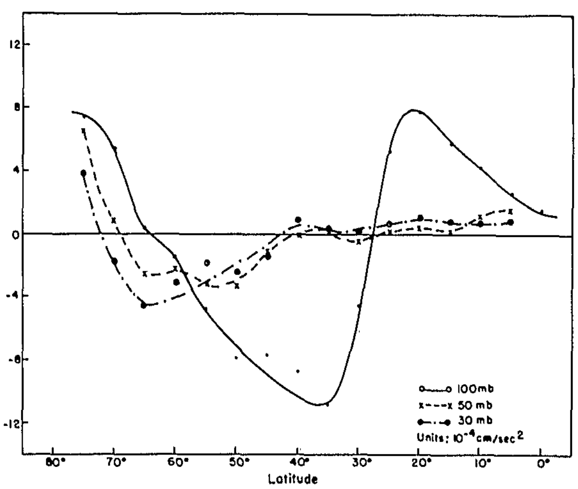

Fig. 1. Horizontal Eddy Momentum Divergence in the Stratosphere Averaged Over One Year, July 57 - June 58, at 100 mb, 50 mb, and 30mb.

$$\frac{\partial \cos^2\phi}{a\cos^2\phi\,\partial\phi}\left([\overline{u'v'}] + [\bar{u}^*\bar{v}^*]\right) \quad \text{in } 10^{-4}\,\text{cm/sec}^2$$

Table 1. Average Horizontal Eddy Momentum Divergence by Seasons.

Summer: July - Sept 57 and April - June 58, 00Z and 12Z.

Winter: Oct - Dec 57 and Jan - March 58, 00Z and 12Z

$$\frac{\partial \cos^2\phi}{a\cos^2\phi\,\partial\phi}\left([\overline{u'v'}]+[\bar{u}^*\ \bar{v}^*]\right) \quad \text{in} \quad cm/sec^2$$

	100 mb		50 mb		30 mb	
Lat.	Summer	Winter	Summer	Winter	Summer	Winter
75	2.3	12.6	.7	12.6	.8	6.9
70	4.0	6.8	.8	1.0	.3	-3.9
65	-2.3	2.1	1.1	-6.2	.1	-9.3
60	.2	-3.1	1.5	-5.6	-1.4	-4.9
55	.3	-9.8	-1.8	-4.2	-1.5	-2.3
50	-3.9	-12.0	-4.6	-1.8	-1.7	-3.1
45	-5.4	-9.9	-2.4	.2	.1	-3.1
40	-5.8	-11.7	1.1	-1.2	2.7	-.7
35	-8.2	-13.6	1.4	-1.2	-.4	1.1
30	-5.4	-3.6	.2	-.9	-.2	.6
25	1.4	9.2	-.3	.7	0.0	1.1
20	3.7	12.0	-1.0	1.9	-.4	2.5
15	4.3	7.4	-.4	1.0	-.3	1.9
10	4.9	3.7	1.5	.7	1.0	.6
5	3.3	2.1	2.1	1.2	1.5	.3

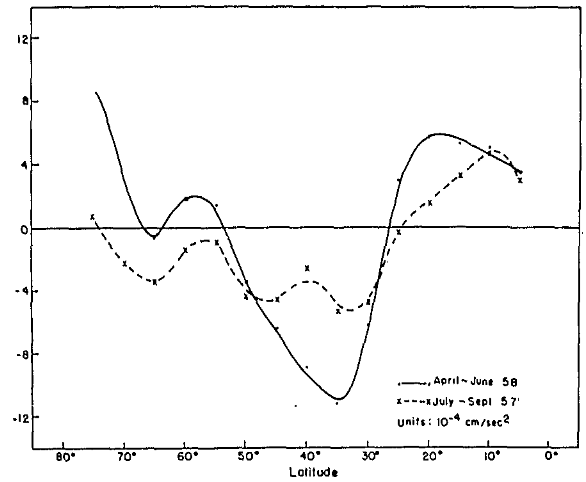

Fig. 2. Horizontal Eddy Momentum Divergence

$\frac{1}{a\cos^2\phi}\frac{\partial}{\partial\phi}\cos^2\phi\left([\overline{u'v'}]+[\bar{u}^*\bar{v}^*]\right)$ at 100 mb. 00Z - 12Z Summer.

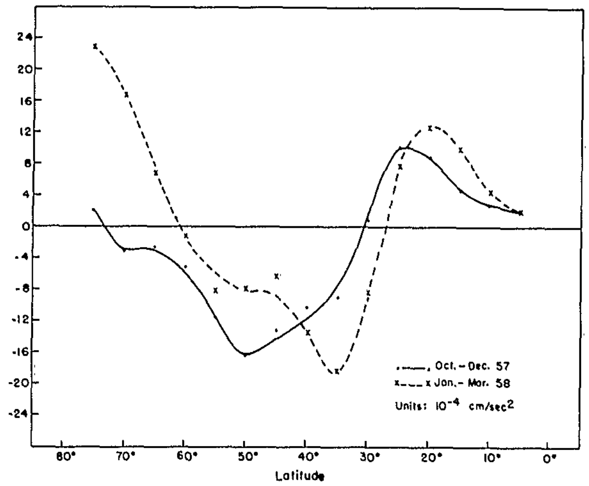

Fig. 3. Horizontal Eddy Momentum Divergence at 100 mb. 00Z - 12Z Winter.

$\frac{1}{a\cos^2\phi}\frac{\partial}{\partial\phi}\cos^2\phi\left([\overline{u'v'}]+[\bar{u}^*\bar{v}^*]\right)$

VII. Momentum Divergence by Vertical Eddies.

The annual mean of the term $\frac{\partial}{\partial p}\left([\overline{\omega' u'}] + [\bar{\omega}^* \bar{u}^*] \right)$ was estimated by finite difference methods from data for January 1958 - June 1958 at 100 mb, 50 mb and 30 mb and could be considered a result for 50 mb or a representative value for the stratosphere. The ω's were first computed by an adiabatic vertical motion equation which is known to give reliable daily values of ω and thus of $\overline{\omega' u'}$. According to Barnes (1962), the mean term $[\bar{\omega}]$ computed by adiabatic methods is incorrect because of the accumulated effect of diabatic heating, but the standing eddies $\bar{\omega}^*$ can be considered reliable. Use is not made of the computed adiabatic $[\bar{\omega}]$, as an estimate of the total $[\bar{\omega}]$ is possible. (See Section VIII).

The results (Figure 5) show a divergence of momentum by vertical eddies over the northern hemisphere south of 60°N, which follows from a downward flux of momentum decreasing with height between level one (about 75 mb) and level two (about 40 mb), at which levels the vertical fluxes were computed. The vertical eddy flux divergence is smaller than the horizontal convergence at 50 mb except where the latter changes sign.

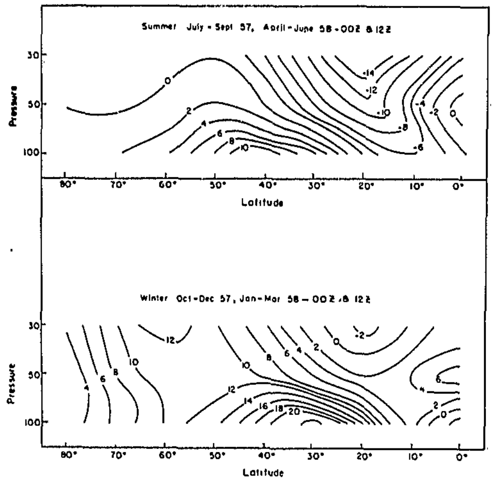

Fig. 4. Mean Momentum per Unit Mass. (Zonal Wind)

$[\bar{u}]$ in meters/sec.

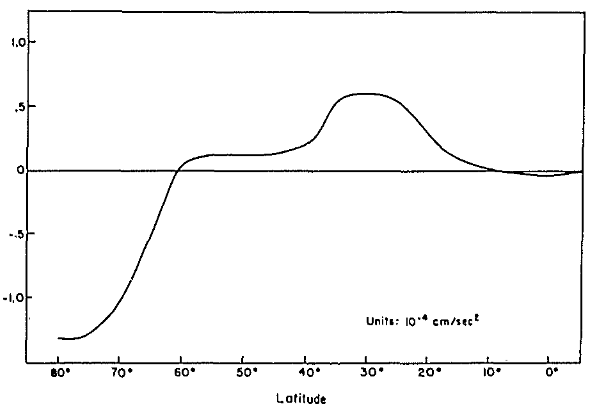

Fig. 5. Momentum Divergence due to Vertical Eddies Jan. - June 58

$$\frac{\partial}{\partial p}\left([\overline{\omega' u'}] + [\bar{\omega}^* \bar{u}^*] \right)$$

VIII. Divergence of Relative Momentum by Mean Motions

The relative momentum convergence due to mean motions has been written $[\bar{\omega}]\frac{\partial[\bar{u}]}{\partial p}+[\bar{v}]\frac{1}{a\cos\phi}\frac{\partial}{\partial\phi}\cos\phi\,[\bar{u}]$

The horizontal and vertical derivatives of the field of u can be accurately computed and are given in figure 6 and figure 7. We note that the term $\frac{1}{a\cos\phi}\frac{\partial}{\partial\phi}\cos\phi\,[\bar{u}]$ is one order of magnitude smaller than the Coriolis parameter f except when within ten degrees of the equator. This would also be true for the troposphere on a mean basis, so mean relative horizontal fluxes of such form will always be negligible compared to fluxes of the earth's angular momentum at any one level.

Mean values of $[\bar{\omega}]$ obtained by taking the average of daily ω adiabatically computed are available for the stratosphere for the IGY, but their validity is questionable because of the neglect of diabatic heating (Barnes 1962). It is possible to measure the $[\bar{\omega}]$ from a mean form of the first law of thermodynamics if the mean heating and eddy fluxes of heat are known. Thus one can write for the momentum convergence due to vertical mean motions.

$$(5)\quad [\bar{\omega}]\frac{\partial[\bar{u}]}{\partial p}=\left(\frac{Q}{C_p}\left(\frac{1000}{p}\right)^{\kappa}+H-[\bar{v}]\frac{\partial[\bar{\theta}]}{a\,\partial\phi}\right)\frac{\frac{\partial[\bar{u}]}{\partial p}}{\frac{\partial[\bar{\theta}]}{\partial p}}$$

where $$H=-\left(\frac{1}{a\cos\phi}\frac{\partial}{\partial\phi}\cos\phi\left([\overline{v'\theta'}]+[\bar{v}^{*}\bar{\theta}^{*}]\right)+\frac{\partial}{\partial p}\left([\overline{\omega'\theta'}]+[\bar{\omega}^{*}\bar{\theta}^{*}]\right)\right)$$

The mean vertical motion can be considered as forced by the product of mean horizontal motions and the latitudinal potential temperature gradient, diabatic heating, and eddy heat fluxes. The resulting momentum transport is the product of the above terms and the vertical wind shear $\frac{\partial[\bar{u}]}{\partial p}$

Consider the term

$$\left(\frac{\frac{\partial[\bar{\theta}]}{a\,\partial\phi} \times \frac{\partial[\bar{u}]}{\partial p}}{\frac{\partial[\bar{\theta}]}{\partial p}} \right) [\bar{V}]$$

In the stratosphere

$$\frac{\partial[\bar{\theta}]}{a\,\partial\phi} < 10^{-7}\ \mathrm{deg/cm}\ ; \qquad \frac{\partial[\bar{\theta}]}{\partial p} > 2\ \mathrm{deg/mb}$$

as computed from data of Murakami (1962).

$$\frac{\partial[\bar{u}]}{\partial p} < 40\ \mathrm{cm/mb\ sec}$$

from Figure 5, thus

$$\frac{\frac{\partial[\bar{\theta}]}{a\,\partial\phi}}{\frac{\partial[\bar{\theta}]}{\partial p}} \cdot \frac{\partial[\bar{u}]}{\partial p} < 2\cdot 10^{-6}\,\mathrm{sec}^{-1} \ll 10^{-4}\,\mathrm{sec}^{-1} \simeq f$$

and so can be neglected compared to momentum convergence due to the Coriolis term.

There is some disagreement among investigators as to the mean diabatic heating in the stratosphere. We used the latest computed results for diabatic heating (Manabe and Moller 1961) plus horizontal eddy transports of temperature

in the stratosphere (Murakami 1962) to estimate the magnitude of the momentum transport due to mean vertical motions. This gives

$$\left| [\overline{\omega}] \frac{\partial [\overline{u}]}{\partial p} \right| \begin{array}{l} \leq 10^{-4} \text{ cm/sec}^2 \text{ at } 100 \text{ mb} \\ \leq \frac{1}{2} \cdot 10^{-4} \text{ cm/sec}^2 \text{ at } 50 \text{ mb and } 30 \text{ mb} \end{array}$$

where the maximum values of middle latitudes are positive. Because vertical heat fluxes were not available at the time of writing, and the mean heating is somewhat uncertain, complete results could not be computed by this method. The estimated momentum convergence values for vertical mean motion are an order of magnitude smaller than horizontal eddy flux convergences. A net annual heating (or cooling) due to eddy heat transport and diabatic heating of one half degree per day at 50 mb or 30 mb, if occurring in middle latitudes where the vertical shear is largest (around 25 cm/sec mb), would give a convergence (divergence) of momentum of about 10^{-4} cm/sec^2. In the troposphere where the static stability is less than in the stratosphere, larger $[\overline{\omega}]$ would result from the same amount of heating as mentioned above.

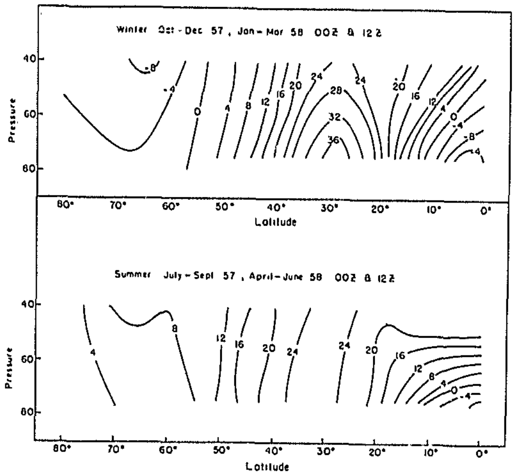

Fig. 6. Zonal Wind Shear $\frac{\partial[\bar{u}]}{\partial p}$ $\frac{cm}{sec\ mb.}$

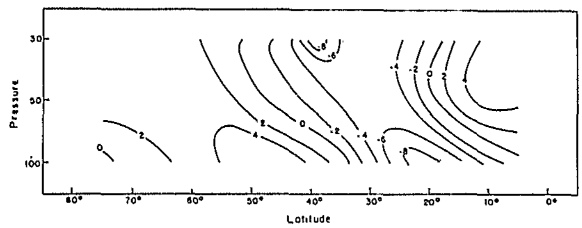

Fig. 7. Annual Zonal Averaged Vorticity in 10^{-6}/sec. $-\frac{\partial[\bar{u}]\cos\phi}{a\cos\phi\,\partial\phi}$

IX. Changes of Zonal Momentum

The term $\overline{\frac{\partial [u]}{\partial t}} = \frac{[u]_{t_2} - [u]_{t_1}}{t_2 - t_1}$ at 100 mb, 50 mb and 30 mb was obtained from geostrophic mean zonal winds computed from some of the IGY stratospheric height maps of the United States Weather Bureau (1960). At the beginning of July 1957 the zonal wind was negative everywhere at 50 mb and 30 mb and in low latitudes at 100 mb. From the time of the summer easterlies the stratosphere gains zonal momentum until it reaches a maximum in January, at which time westerly winds prevail throughout the stratosphere. Easterlies are again found at the end of the period studied, July 1958. Some changes occur suddenly such as the breakup of the polar night jet, which is described by Teweles (1958), but in any case the large gain of momentum between July and January and the subsequent loss of equal magnitude must be balanced by other terms in equation (4).

In polar latitudes the measured mean eddy fluxes are sufficiently great at all levels so that they must be accounted for in any theory of the dynamics of these regions, but in the stratosphere at 50 and 30 mb southward of 30°N the annual mean eddy flux divergences are quite small. If the seasonal momentum changes of the equatorial stratosphere were to be balanced by eddy momentum divergences, the eddy divergence of the period January - June 1958 would have to be greater than the annual mean value. From figure 8 and figure 9 it is seen that this is not sufficiently so at 50 and 30 mb. Thus, synoptic scale eddy fluxes in

this region might be neglected in a rough theoretical formulation of the seasonal changes of the zonal wind at these levels. The smallness of the eddy flux divergences in the middle stratosphere is a rather unique result. These divergences are much smaller than the seasonal changes of momentum in the troposphere and at 100 mb in the lower stratosphere. Such simpler behavior in the middle equatorial stratosphere is perhaps analogous to the Hadley regimes of rotating tank experiments.

If a time period is centered about winter or summer, or if it covers a period of a year, the term $\overline{\frac{\partial [u]}{\partial t}}$ will be much smaller and the eddy fluxes become again the largest measured term. $\overline{\frac{\partial [u]}{\partial t}}$ For July 1957 - June 1958 in the stratosphere was found $\leq .1 \times 10^{-4}$ cm/sec^2, which is an order of magnitude smaller than the measured flux convergences. An exception is the biannual zonal momentum change characteristic of the equatorial stratosphere wind oscillation, as described by Reed and Rogers (1961). It is interesting to see that the computed horizontal eddy momentum divergence, positive in the immediate vicinity of the equator, is of the proper sign at 50 mb and 30 mb to account for the negative $\overline{\frac{\partial [u]}{\partial t}}$ during the time of our study. According to the time-height cross section of Reed and Rogers (1961, page 128) for Canton Island $\overline{\frac{\partial [u]}{\partial t}} = -1.2 \cdot 10^{-4}$ cm/sec^2 at 30 mb. In the present study a mean eddy momentum divergence of $0.9 \cdot 10^{-4}$ cm.sec^{-2} for 5°N was computed at the same level.

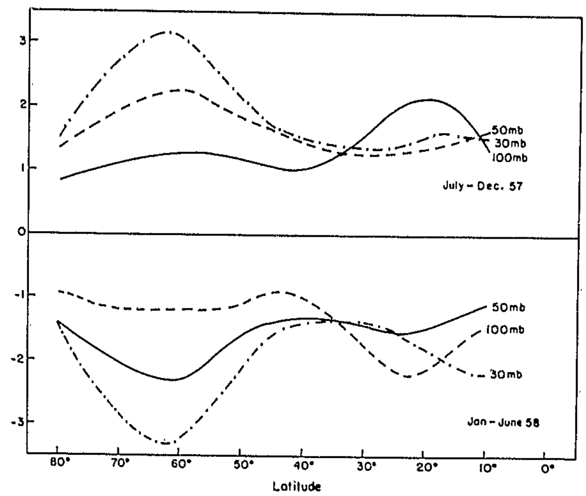

Fig. 8. Mean Zonal Acceleration $\frac{\overline{\partial[\bar{u}]}}{\partial t}$ 10^{-4} cm sec^{-2}.

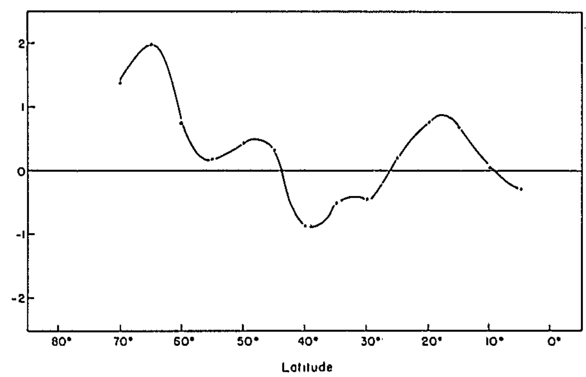

Fig. 9. Momentum Divergence Deviation from Annual Average Jan. - June 58 Minus Annual Mean (July 57 - June 58) Average of 50 mb and 30 mb. 10^{-4} cm.sec^{-2}.

X. Friction and Zonal Wind Curvature

The barotropic behaviour of baroclinically developing eddies of synoptic scale results in an upgradient momentum transport in the troposphere and also in regions of the stratosphere as evident from the data of this study. However, the concept of eddy viscosity may still be applicable to small-scale turbulence. If one assumes small-scale eddies are present which transfer momentum as does Newtonian viscosity, the momentum divergences due to eddy friction can be computed from the curvature of the zonal wind profiles. Vertical wind shear is much larger than horizontal shear, so momentum transfer according to eddy viscosity theory should produce momentum divergences as $-K \frac{\partial^2 [\bar{u}]}{\partial z^2}$. From the 100mb, 50 mb and 30 mb data, finite difference values for $\frac{\partial^2 [\bar{u}]}{\partial z^2}$ at 50 mb for summer and winter were obtained (see figure 10). For both time periods $\frac{\partial^2 [\bar{u}]}{\partial z^2}$ is significantly positive between 20° and 60°N, which, according to the assumptions of eddy viscosity, can only result in a gain of momentum. Where the jet is a maximum in the vicinity of 200 mb, $\frac{\partial [\bar{u}]}{\partial z} = 0$ and from viscosity theory it follows that there can be no small-scale eddy momentum transported across the jet peak. The curvature of the mean zonal wind sufficiently near the jet is expected to be negative and to vanish somewhere in the vicinity of 100 mb. An estimate of the annual mean curvature at 100 mb using mean 200-mb data from the United States Weather Bureau cross sections (1961) suggests a curvature at 100 mb in middle latitudes much smaller than at 50 mb. It is impossible to assign a numerical value to friction because of ignorance of the eddy viscosity

coefficient K. We can, however, compute a maximum value by certain assumptions. The only measured momentum source for the region immediately above the tropospheric jet is horizontal eddy convergences of momentum. A poleward meridional circulation would also be a momentum source, but previous data studies of $[\bar{v}]$ suggest that if anything the meridional circulations in the regions of maximum westerly wind are negative (Starr 1954). We assume that one half of the momentum which is known to converge in middle latitudes by horizontal eddy momentum fluxes between 200 mb and 100 mb from previous data studies such as Starr and White (1954) and Buch (1954) is transported above this region into the middle stratosphere. We assume that our measured curvature at 50 mb is a maximum and that $\frac{\partial^2[\bar{u}]}{\partial z^2}$ decreases linearly to zero at 100 mb and 10mb. Then $K = 1/2 \cdot 10^6$ cm^2/sec in the stratosphere. Haurwitz (1961), in discussing possible frictionally driven mesospheric meridional circulations, suggested a K twenty times as large to be reasonable. On the other hand, there is little experimental evidence for viscosity above the earth's friction layer, so K may be much smaller than the chosen value.

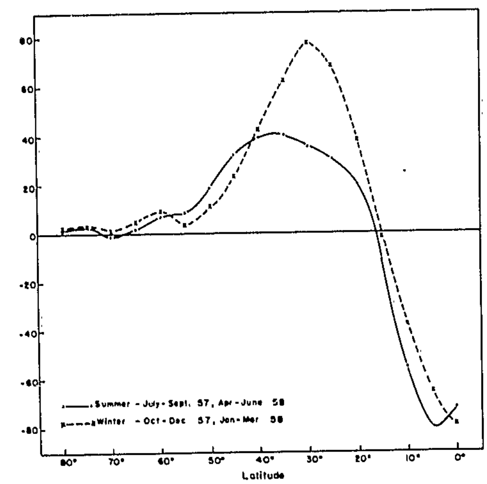

Fig. 10. $\frac{\partial^2 [\bar{u}]}{\partial z^2}$ Curvature of Mean Zonal Wind in 10^{-10} cm/sec at 50 mb.

XI. Concluding Remarks

Because of the great difficulty of obtaining statistically significant mean meridional circulation, $[\bar{v}]$, there is some value in computing them from indirect measurements of momentum divergences. Doubt concerning the magnitude of the vertical momentum transport by small-scale eddies is a limitation on such computations. Direct wind measurements will be discussed in later publications by other members of the Planetary Circulations Project staff.

From equation (4) one gets

$$[\bar{v}] = \left(f - \frac{1}{a\cos\phi}\frac{\partial}{\partial\phi}\cos\phi[\bar{u}]\right)^{-1}\left(\overline{\frac{\partial[u]}{\partial t}} + \frac{1}{a\cos^2\phi}\frac{\partial}{\partial\phi}\cos^2\phi\left([\overline{u'v'}]+[\bar{u}^*\bar{v}^*]\right)\right.$$
$$\left. + \frac{\partial}{\partial p}\left([\overline{u'\omega'}] + [\bar{u}^*\bar{\omega}^*]\right) + [\bar{\omega}]\frac{\partial[\bar{u}]}{\partial p} - \text{Friction}\right) \tag{6}$$

which is exact.

All of the terms have been measured except friction. The actual $[\bar{v}]$ can be considered as composed of a linear sum of contributions on the right-hand side of (6).

In this study we have computed all terms of importance for indirectly measuring the $[\bar{v}]$ except friction. Over an annual period $\overline{\frac{\partial[u]}{\partial t}}$ can be neglected and the horizontal fluxes are the most significant measured quantity. In figure 11 is given the $[\bar{v}]$ at 100 mb due to horizontal eddy divergences for the year of this study. In figure 12 is plotted $[\bar{v}]$ (annual mean) due to horizontal eddy divergences plus vertical

eddy divergences at 50 mb and also the $[\bar{v}]$ resulting from these plus the maximum possible eddy viscosity surmised in Section X. If friction is negligible in the stratosphere, one can deduce from the results of this study that there exists at 100 mb an indirect circulation in middle latitudes north of 28°N of greater strength in winter than in summer with an annual average value of around 10 cm/sec, a direct circulation of order of 20 cm/sec being found near the equator. The frictionless meridional circulation for a one-year average at 50 or 30 mb is only a few centimeters per second, an indirect circulation in polar latitudes being the most significant feature. The seasonal changes of zonal momentum discussed in Section IX are the largest measured term of equation (4) in the equatorial stratosphere at 50 mb and 30 mb so the seasonal frictionless meridional circulation in equatorial regions in comprised largely of an oscillation which is poleward in fall and equatorward in spring with velocities of about five cm/sec. In figure 12, the circulations deduced from a maximum eddy viscosity show a strong equatorward motion in middle latitudes. Haurwitz (1961) has shown independently that eddy friction up to 40 km implies equatorward motion in middle latitudes. Thus there is no known term in (6) to give a northward motion in middle latitudes. This confirms theories of northward eddy transport of ozone and radioactive materials as discussed by Newell (1961).

It is useful to summarize the orders of magnitude of the various terms in equation (4) as far as it is possible to deduce them from

this study. This is done in Table 2. Such experimental scale analyses of the known equations governing the general circulation statistics, together with experimental and theoretical efforts to find new relationships governing these statistics are necessary to make progress in dynamical climatology; which latter is defined as the seeking of a complete set of equations governing the statistics of the atmosphere over various time scales.

The directly measured southward $[\bar{v}]$ of Oort (1962) are almost an order of magnitude larger than the $[\bar{v}]$ forced by horizontal eddy transports. They would produce a rapid rate of decrease of zonal angular momentum unless the stratosphere were dragged around by friction as suggested by Starr (1960). This may be the case unless some systematic error is present in the $[\bar{v}]$ measured from actual wind data. The semidiurnal oscillation of wave number zero (resolved at the ground by pressure oscillation data on a global basis) may be sufficiently large in the stratosphere to result in such a large systematic error. For instance, data in the Azores presented by Harris, Finger and Teweles (1962) shows semidiurnal oscillations of the order of a half a meter per second. Without information on these oscillations sufficient to define the variation and mean around a latitude circle, no conclusions can be drawn.

Table II. Order of Magnitude of Various Terms in Equation (4) in 10^{-4} cm/sec^2.

$\overline{\frac{\partial [u]}{\partial t}}$	Summer to Winter	100 mb	2 Eq. Strat.	1 Polar Strat.
		50, 30 mb	2 Eq. Strat.	3 Polar Strat.
	Annual Value	$\leq$ .1 except in vicinity of equator where it is $-1 \leq \frac{\partial [\bar{u}]}{\partial t} \leq 1$ at 50, 30 mb depending on the phase of the 26 m cycle.		
	Annual Values			
$\frac{1}{a \cos^2\phi} \frac{\partial}{\partial \phi} \cos^2\phi \left([\overline{u'v'}] + [\bar{u}^* \bar{v}^*] \right)$		100 mb	$\sim$ 10	
		50, 30 mb	$\sim$ 1 Eq. Strat. $\sim$ 3 Polar Strat.	
$\frac{\partial}{\partial p} \left([\overline{u'\omega'}] + [\bar{u}^* \bar{\omega}^*] \right)$		100 mb	$\sim$ 1 estimated	
		50 mb	$\sim$ 1/2	
		30 mb	$\sim$ 1/2 estimated	
$[\bar{\omega}] \frac{\partial [\bar{u}]}{\partial p}$		100 mb	$\sim .5 \cdot 10^{-4}$	
		50, 30 mb	$\sim .25 \cdot 10^{-4}$	
$f [\bar{v}]$		+ friction $\cong$	horizontal eddy divergence over an annual period	

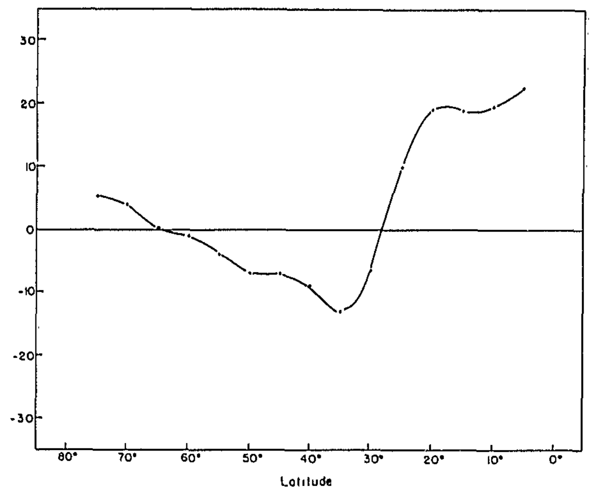

Fig. 11. Indirectly Measured Mean Meridional Circulation Due to Horizontal Eddy Fluxes. $[\bar{v}]$ in cm/sec

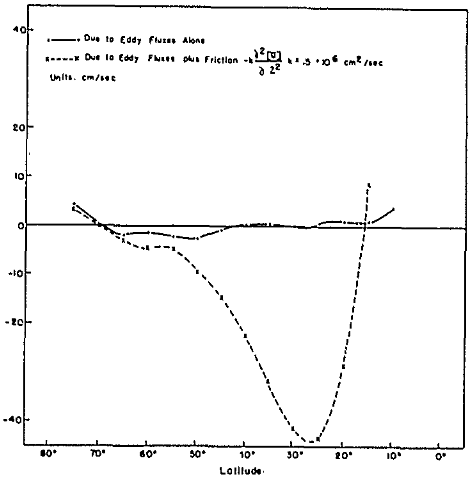

Fig. 12. Indirectly Measured Mean Meridional Circulation at 50 mb. $[\bar{v}]$ in cm/sec

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Direct Measurement of the Meridional Circulation in the Stratosphere during the IGY

by

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ABSTRACT

The horizontal component of the meridional circulation at the 100, 50 and 30 mb levels in the stratosphere is measured for the IGY period July 1957 through June 1958. Data from approximately 240 stations, well distributed over the northern hemisphere, are used in the hemispheric analysis of the north-south component of the wind. A separation of the data into four periods of three months enabled us to determine seasonal variations. In addition, 00Z and 12Z data were treated separately in order to detect possible diurnal effects. Values of the mean meridional circulation, that is the zonal average of the time mean north-south component of the wind, are presented at every 5° latitude from the equator to 80°N.

In the yearly average at 100 mb an "apparent" three cell pattern is observed, consisting of a region of strong poleward motion north of 55°N, equatorward motion from 15° to 55°N and weak poleward motion from 0° to 15°N. In the yearly averages at 50 and 30 mb poleward motions appear only at high latitudes and diminish in strength with height. Equatorward motions dominate at 30 mb. A maximum value of the mean meridional circulation of + 60 cm/sec is found at 100 mb at 70°N.

A comparison is drawn with the meridional circulation in the stratosphere of the southern hemisphere, measured by Obasi (1963). This circulation forms practically a mirror image of the northern hemispheric circulation.

The dominant equatorward character of the mean flow found at 30 mb is compared with the mean rimward velocities observed by Starr and Long (1953) at the top of the fluid in a rotating tank experiment.

1. Introduction

Research on the general circulation of the earth's atmosphere generally has been directed at a description of the behaviour of the atmosphere in a simple way. The differential heating of the atmosphere (relative warming near the equator and cooling near the poles) suggests a strong direct circulation with rising of warm air at the equator and sinking of cold air in the polar regions. However, because of the rotation of the earth the actual wind systems present in the troposphere show a system of direct Hadley cells at polar and equatorial latitudes and an indirect Ferrel cell in middle latitudes, all of which are rather weak and difficult to measure. The indirect circulation implies a southward flow at the levels of the tropospheric jet stream. This southward flow finally indicates a large difficulty with the formerly common hypothesis of a mean meridional circulation as the source of angular momentum and energy for the strong zonal wind systems. The organized horizontal eddy motions, instead of a simple toroidal overturning are now known to supply sufficient energy and momentum for the maintenance of the jet stream (Starr and White, 1951).

It is of considerable interest to measure the intensity and direction of these meridional circulations in the higher layers of the atmosphere (at altitudes of 16-24 km). Prior to the IGY the observational material for the stratosphere was sparse, so that it was not possible at that time

to obtain from actual wind measurements the nature of the meridional circulations. Murgatroyd and Singleton (1961) attempted to derive the mean meridional circulations indirectly. They assumed a certain distribution of heating and cooling with latitude and height and neglected effects of eddies. In the present study direct wind measurements are used to determine the mean circulation at the 100, 50 and 30 mb levels. It is shown that the circulations based on actual wind observations are quite different from the simplified picture presented by Murgatroyd and Singleton.

A dense network of 240 stations over the northern hemisphere was chosen. A year of data (July 1957 - June 1958) gave a smooth pattern, consistent at all levels, for the mean meridional circulations.

In the last section of this paper we will compare our results with the circulations in the southern hemisphere, as determined by Obasi (1963) during an overlapping period of the IGY, and with the meridional velocities in a rotating tank experiment, as observed by Starr and Long (1953).

2. DATA AND THEIR REPRESENTATIVENESS

The data used in this study are those reported on the IGY microcards, issued by the World Meteorological Organization in Geneva. The stations were selected on the basis of good observations of winds and temperatures up to the stratospheric levels 100, 50 and 30 mb. Except in the eastern Pacific from 100^{o}W to 140^{o}W at low latitudes, the coverage over the northern hemisphere is very good and far superior to that employed in similar studies prior to the IGY. A total number of 240 stations was used (for the geographical location of the stations see Oort, 1962).

For each station and each level three months of wind observations were averaged in order to obtain time mean values of the horizontal meridional wind component. The necessary computations were performed with a Royal-Mc Bee LGP-30 computer. The periods considered are July-Sept 1957, Oct-Dec 1957, Jan-March 1958, and April-June 1958; it is thought that the averages over these periods give a representative picture for a season.

The computed values of $\overline{V}$ were plotted on polar stereographic maps, which were then analyzed by hand. Maps for 00Z and 12Z were analyzed separately so that it would be possible to get an idea of the diurnal effects and also to detect computational errors in the plotted station values. Since observations at 00Z and

12Z were missing for the African stations south of 15°N, 06Z and 18Z data were used for the analysis in the region over Africa. As in Barnes' (1963) and Murakami's study (1962) little weight was given to stations with less than 30 observations out of a total possible 90. From the analyzed maps grid point values were read at every 5° latitude from the equator to 80°N and at every 10° longitude, throughout the hemisphere. Zonal averages were computed as the average of the 36 longitude grid point values along a latitude circle. The maps for the periods July-Sept 1957 and Oct-Dec 1957, first analyzed by Murakami (1962), were reanalyzed with the help of additional information. The total amount of data in this study covers a period of one year (July 1957 through June 1958).

3. NOTATION

(u, v, ω) = "velocity" components in (x, y, p) coordinate system

u = west-east component of the wind (positive if from the west)

v = north-south component of the wind (positive if from the south)

ω $\equiv$ $\frac{dp}{dt}$ ≡ "vertical velocity"

ϕ = latitude

λ = longitude

g = acceleration due to gravity

f = Coriolis parameter

z = height of isobaric level

$\overline{v} \equiv \frac{1}{t_2 - t_1} \int_{t_1}^{t_2} v \, dt$ = time average of meridional component of the wind

$[v] \equiv \frac{1}{2\pi} \int_{0}^{2\pi} v \, d\lambda$ = zonal average of meridional component of the wind

$[\overline{v}] \equiv \frac{1}{2\pi(t_2 - t_1)} \int_{0}^{2\pi} \int_{t_1}^{t_2} v \, dt \, d\lambda$ = zonal average of the time mean meridional component of the wind ≡ "mean meridional circulation"

4. RESULTS

4.1 Discussion of the observed meridional circulations

The $\overline{v}$ maps drawn for 00Z and 12Z separately show in general good agreement. In a later section of this paper we will come back to the question of possible tidal effects in $[\overline{v}]$. We will now comment on some of the features of the $\overline{v}$ pattern (for presentation of the $\overline{v}$ maps and a more detailed discussion see Oort, 1962):

1. The $\overline{v}$ field is organized in a number of cells around a latitude circle. The cells have a north-south extension of approximately 30° to 40° latitude. Along the same latitude circle we find maxima of $\overline{v}$ of +5 to+10 m/sec and minima of -5 to -10 m/sec.

2. The meridional circulations are strongest at high latitudes and during the winter season. These facts appear to be related to the meandering of the polar jet. Extremes of $\overline{v}$ of the order of 10 m/sec occur at 70° - 80°N during the Jan-March 1958 period.

3. The cells are in general weaker at 50 and 30 mb than at 100 mb (especially in the summer season).

4. North-south "troughlines" in the flow are found at the east sides of the Eurasian and American continents, and over western Europe (the latter is weak).

5. In all four seasons there is a marked reversal of strong southerlies at 100 mb to northerlies at 50 and 30 mb over north

west Africa. A less pronounced reversal from northerlies at 100 mb to southerlies at 50 and 30 mb occurs on the other side of the hemisphere at 10° - 20°N in the vicinity of the international date line.

The areas of positive and negative $\overline{V}$ in a horizontal plane reduce the zonal average $[\overline{V}]$ by more than an order of magnitude. The standard deviation of $\overline{V}$ from its average along a latitude circle, gives a good indication of the cellular pattern (see Table 2).

It should be noticed that it is not possible to use geostrophic winds to compute $[\overline{V}]$, in the (x, y, p) - system, since the zonal average of $V_{geostr.} = -\frac{g}{f}\frac{\partial z}{\partial \lambda}$ vanishes. Therefore isobaric height maps cannot supply information about the mean meridional motions. The only direct method is by careful analysis of actual wind data.

Table 1 gives $[\overline{V}]$ computed for July 1957 - June 1958. We shall now point out the main features of Figures 1, 2 and 3 representing $[\overline{V}]$ for summer and winter at the three levels considered.

<u>100 mb; winter N. H. (Figure 1)</u>

Strong morthward motion poleward of 50°N (maximum of 110 cm/sec at 70°N). Southward motion from 25°N to 50°N (maximum of -60 cm/sec at 35°N). Weak northward motion from the equator to 25°N (about 15 cm/sec). The winter situation at 100 mb resembles the circulation in the upper <u>troposphere</u>, where we seem to have northward motion in

high latitudes (direct cell), southward motion in middle (indirect cell) and again poleward motion in low latitudes (direct cell). However, the northward motion at low latitudes disappears during the summer season.

100 mb; summer N.H. (Figure 1)

Weak northward motion poleward of 60°N (maximum of 30 cm/sec at 75°N). Southward motion from the equator to 60°N (maximum of -40 cm sec at 30°N).

50 mb; winter N.H. (Figure 2)

Northward motion north of 60°N (maximum of 50 cm/sec at 75°N). Southward motion from the equator to 60°N (maximum of -25 cm/sec at 45°N and - 20 cm/sec at 10°N).

50 mb; summer N.H. (Figure 2)

The circulation is similar to that during the winter. The maximum strength of northward motion is 40 cm/sec. A maximum of southward motion is found at 40°N (-25 cm/sec).

30 mb; winter N.H. (Figure 3)

Poleward motion north of 65°N (maximum of 30 cm/sec at 75°N). The midlatitude equatorward velocities are greater than at 50 mb. A

Table 1a. Mean meridional velocity $[\overline{V}]$ at 100 mb obtained from 00Z and 12Z observations for the period July 1957 - June 1958*. Units are cm/sec.

	July-Sept 1957		Oct-Dec 1957		Jan-March 1958		April-June 1958	
	00Z	12Z	00Z	12Z	00Z	12Z	00Z	12Z
80°N	77	48	67	63	52	100	20	-4
75°N	38	0	86	83	59	134	44	14
70°N	16	-30	104	122	78	134	61	28
65°N	14	2	112	131	63	102	41	17
60°N	-2	9	61	106	16	31	8	-8
55°N	-2	12	24	18	2	-14	-17	-24
50°N	-4	12	13	-36	22	0	-27	-28
45°N	8	-9	-13	-49	16	-32	-21	-62
40°N	-22	-13	-46	-58	-33	-87	-32	-79
35°N	-21	-30	-31	-35	-58	-117	-34	-67
30°N	9	-43	-5	-15	-56	-105	-28	-48
25°N	30	-56	32	18	-34	-33	-60	-60
20°N	24	-44	18	6	6	34	-60	-47
15°N	36	-27	-14	-8	33	38	-40	-25
10°N	37	-11	-14	-11	32	32	-27	-8
5°N	27	-8	11	-9	26	25	-18	-3
0°	21	-21	28	-12	19	22	-14	-1

*The data for July-December 1957 are obtained from reanalyzed maps, which were originally analyzed by Murakami (1962).

Table 1b. Mean meridional velocity $[\bar{v}]$ at 50 mb obtained from 00Z and 12Z observations for the period July 1957 - June 1958*. Units are cm/sec.

	July-Sept 1957		Oct-Dec 1957		Jan-March 1958		April-June 1958	
	00Z	12Z	00Z	12Z	00Z	12Z	00Z	12Z
80°N	64	34	-21	19	87	84	37	12
75°N	47	21	13	12	84	77	68	31
70°N	35	11	49	21	66	68	51	17
65°N	3	1	48	21	9	26	-4	-4
60°N	-9	-7	18	-12	-22	-14	-29	-27
55°N	-1	0	-34	-5	-7	-37	-16	-25
50°N	0	7	-60	-7	4	-23	1	-11
45°N	-9	-2	-39	-31	5	-29	-24	-25
40°N	-14	0	-29	-39	6	-37	-52	-32
35°N	-21	4	-3	-30	-1	-49	-67	-36
30°N	-11	21	-3	-17	9	-26	-61	-14
25°N	-5	29	-12	10	-2	-9	-50	-2
20°N	7	26	-17	22	-26	-1	-29	-1
15°N	23	23	-14	18	-43	-12	-29	-15
10°N	28	19	-9	4	-47	-32	-11	-27
5°N	18	9	1	-8	-29	-32	4	-29
0°N	10	5	13	-12	-7	-11	13	-7

*see footnote Table 1a.

Table 1c. Mean meridional velocity $[\bar{v}]$ at 30 mb obtained from 00Z and 12Z observations for the period July 1957 - June 1958*. Units are cm/sec.

	July-Sept 1957		Oct-Dec 1957		Jan-March 1958		April-June 1958	
	00Z	12Z	00Z	12Z	00Z	12Z	00Z	12Z
$80^{o}N$	41	15	62	19	-6	22	-7	-33
$75^{o}N$	22	21	84	64	3	-21	-1	-48
$70^{o}N$	-14	9	64	73	-43	-26	-3	-42
$65^{o}N$	-23	-9	24	47	-45	11	-5	-38
$60^{o}N$	-16	-20	-32	-14	-31	34	-12	-39
$55^{o}N$	1	-19	-90	-68	-34	24	-21	-20
$50^{o}N$	3	-16	-81	-66	-42	-28	-19	-9
$45^{o}N$	-9	-3	-36	-43	-47	-52	-31	-13
$40^{o}N$	-1	1	-6	-19	-62	-60	-47	-25
$35^{o}N$	9	4	16	-9	-59	-44	-35	-17
$30^{o}N$	4	4	24	11	-39	-21	-23	-10
$25^{o}N$	-13	-6	31	11	-18	-21	-4	-11
$20^{o}N$	-31	-4	33	1	-2	-38	8	-16
$15^{o}N$	-18	23	18	-6	-14	-42	9	-25
$10^{o}N$	-4	29	4	-9	-29	-35	9	-24
$5^{o}N$	-1	14	-11	-21	-35	-21	2	-21
0^{o}	2	4	-11	-14	-29	-1	2	-17

*see footnote Table 1a.

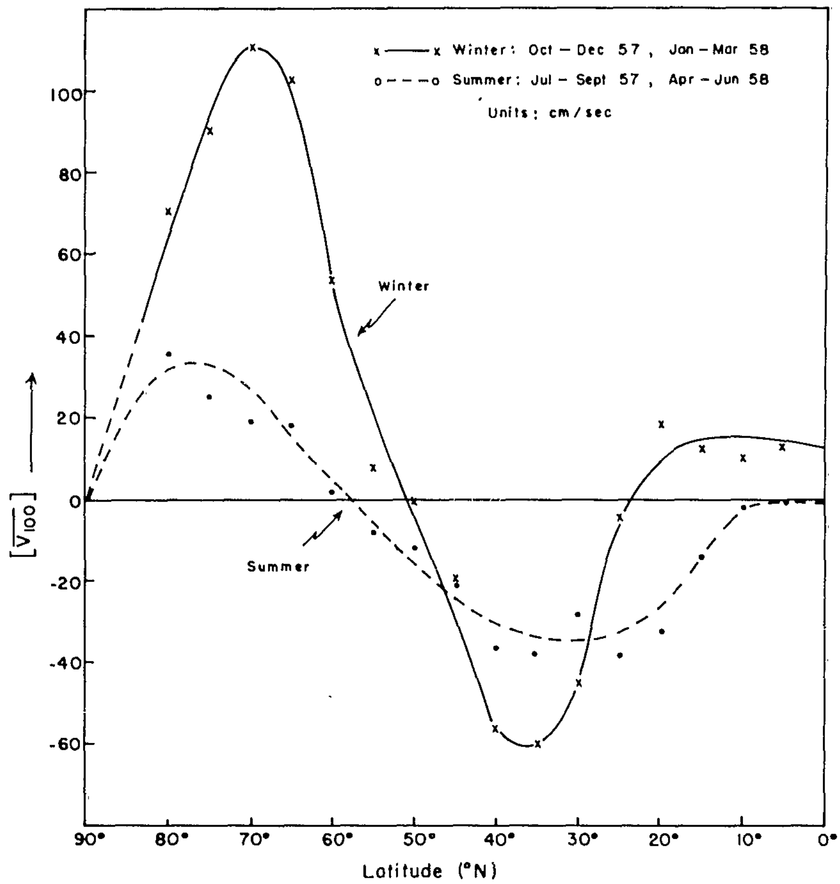

Figure 1. $[\overline{V}]$ at 100 mb for winter and summer 1957/1958. Positive $[\overline{V}]$ indicates northward motion, negative $[\overline{V}]$ indicates southward motion. Units are cm/sec. 00Z and 12Z data are averaged.

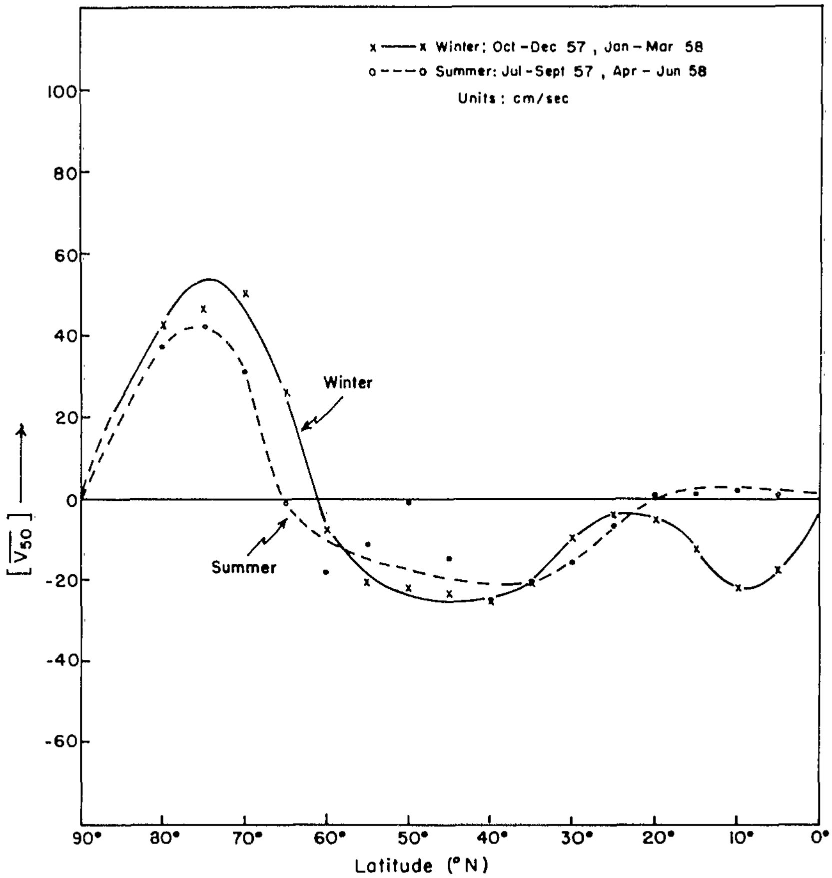

Figure 2. [$\overline{V}$] at 50 mb for winter and summer 1957/1958. Positive [$\overline{V}$] indicates northward motion, negative [$\overline{V}$] indicates southward motion. Units are cm/sec. 00Z and 12Z data are averaged.

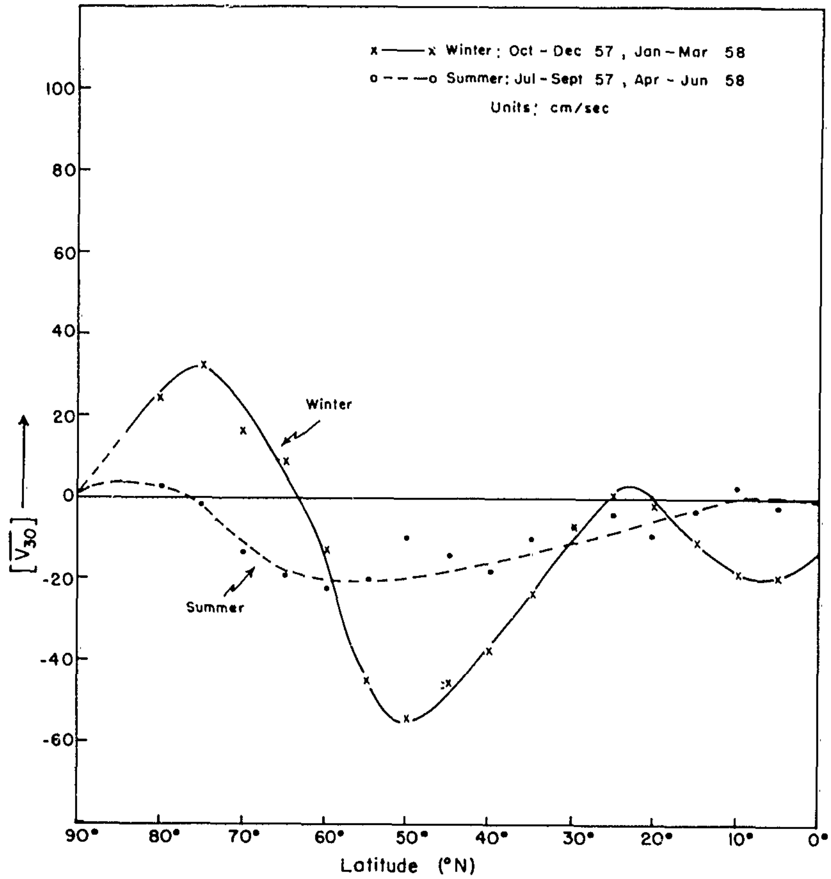

Figure 3. $[\overline{V}]$ at 30 mb for winter and summer 1957/1958. Positive $[\overline{V}]$ indicates northward motion, negative $[\overline{V}]$ indicates southward motion. Units are cm/sec. 00Z and 12Z data are averaged.

Table 2. The standard deviation $\sigma(\overline{v})$ of $\overline{v}$ from its zonal average $[\overline{v}]$ for January-March 1958 and April-June 1958. 00Z and 12Z observations are averaged. Units are m/sec.

	Jan-March 1958			April-June 1958		
	100 mb	50 mb	30 mb	100 mb	50 mb	30 mb
80° N	6.72	7.90	6.22	1.93	1.74	1.26
75°	6.94	8.26	6.22	2.74	2.52	1.34
70°	6.39	7.43	6.36	2.89	2.36	1.48
65°	5.56	5.80	5.84	2.52	1.78	1.59
60°	5.17	5.12	5.09	2.29	1.61	1.56
55°	4.99	4.57	4.56	2.18	1.61	1.52
50°	4.32	4.13	4.11	2.18	1.64	1.49
45°	3.58	3.58	3.27	2.32	1.62	1.26
40°	2.79	2.76	2.26	2.63	1.57	0.97
35°	2.32	1.97	1.60	2.78	1.44	0.81
30°	2.76	1.48	1.18	2.85	1.11	0.68
25°	3.18	1.26	1.10	2.62	0.76	0.66
20°	3.17	1.00	0.93	2.26	0.64	0.72
15°	2.96	1.06	0.81	1.90	0.74	0.79
10°	2.56	1.10	0.69	1.46	0.74	0.62
5°	1.92	0.93	0.64	1.02	0.63	0.46
0°	1.33	0.68	0.48	0.68	0.46	0.30

maximum of -55 cm/sec is found at 50°N.

30 mb; summer N. H. (Figure 3)

Practically no mean poleward motion is in evidence. A broad maximum of - 20 cm/sec is centered around 55°N.

Seasonal cross sections

Cross sections of $[\overline{v}]$ for the four three month periods are presented in Figures 4 and 5. A significant change in $[\overline{v}]$ takes place near the equator in the course of 1957/1958. Mean equatorward motions seem to replace the poleward motions at low latitudes, which were present in the first period July-Sept 1957. It is possible that this gradual change in $[\overline{v}]$ is related to the equatorial 26 month wind cycle. More than one year of data are needed to draw more definite conclusions.

Implications for the transport of ozone

The January through June cross sections show that the mean meridional transport of mass at these levels is towards the equator during the spring of 1958 (from 0° to 65°N). It is often assumed that mean meridional motions transport ozone from the equator towards high latitudes during the spring. This assumption is made in order to

account for the spring build up of ozone in middle and high latitudes. Our data show that this explanation of a mean meridional circulation as a mechanism for the northward ozone transport is not correct. The $[\overline{v}]$ work in a direction of accumulating ozone in the equatorial regions. Newell (1961) estimates that horizontal eddy processes can give the required northward transport during the spring (see also Martin, 1956).

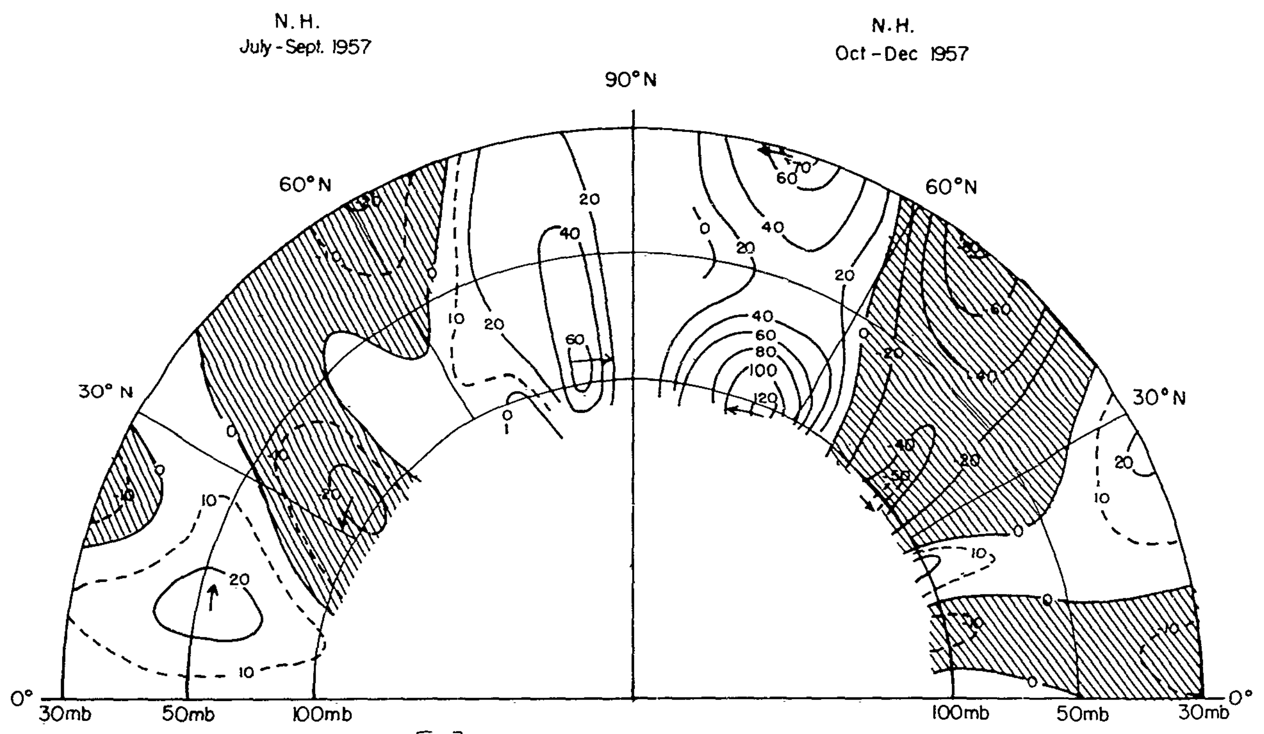

Figure 4. Cross section of $[\bar{v}]$ through the stratosphere (100 - 30 mb) of the northern hemisphere for the periods July - Sept 1957 and Oct - Dec 1957. 00Z and 12Z observations are averaged. The regions of equatorward motion are shaded. Units are cm/sec.

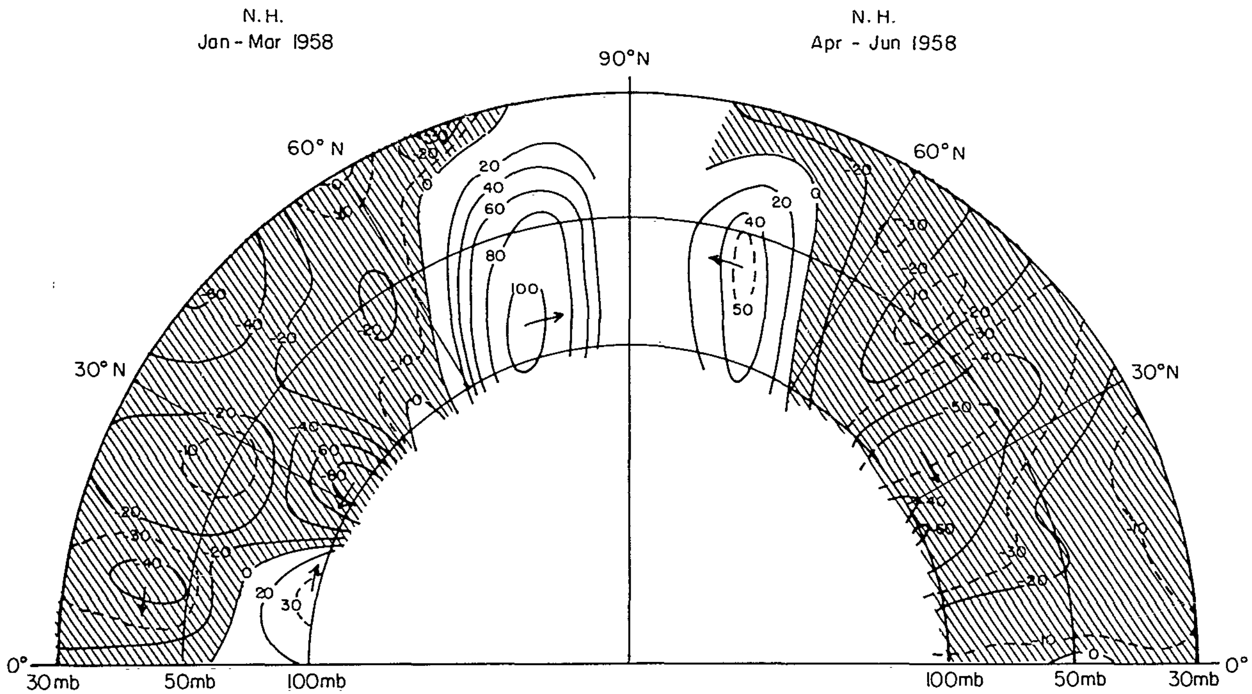

Figure 5. Cross section of $[\overline{v}]$ through the stratosphere (100 - 30 mb) of the northern hemisphere for the periods Jan - March 1958 and April - June 1958. 00Z and 12Z observations are averaged. The regions of equatorward motion are shaded. Units are cm/sec.

4.2 Discussion of diurnal effects in the mean meridional circulation

The mean meridional motions, presented in Table 1 show some systematic differences in the data for 00Z and 12Z. These differences are partly due to sampling errors, but also to real diurnal variations in the winds. In order to measure representative mean meridional circulations in the atmosphere it would be necessary to sample the wind field at least four times a day e.g. at 00Z, 06Z, 12Z and 18Z. The average of the data for the four standard hours would remove the most dominating diurnal and semidiurnal effects. An evident shortcoming of our study is that semidiurnal effects may still be important. Other members of the M.I.T. Planetary Circulations Project are looking into the possibility of an extension of this study to include also 06Z and 18Z data. However, a handicap of such a project is the relative scarcity of the observations at 06Z and 18Z over most parts of the hemisphere.

In Table 3 we present the diurnal differences of the mean meridional circulations for winter and summer 1957/1958. These so called "tidal" effects in the atmosphere seem to be the result of heating (see e.g. Siebert, 1961). Because of changes in the declination of the sun we do not present in this table the yearly average, but only the winter and summer averages. A significant feature which shows up at 50 mb in all seasons appears to be the relatively stronger southward motions at 00Z than at 12Z (the difference is approximately 30 cm/sec) around 30°N in the summer and around 15°N in the winter.

Table 3. Summer and winter averages of the diurnal variation of the mean meridional circulation at 100, 50 and 30 mb for the northern hemisphere. Units are cm/sec.

	Summer (July-Sept '57 + April-June '58) $[\bar{v}]_{00z} - [\bar{v}]_{12z}$			Winter (Oct '57 - March '58) $[\bar{v}]_{00z} - [\bar{v}]_{12z}$		
	100mb	50mb	30mb	100mb	50mb	30mb
80°N	27	27	26	-22	-18	7
75°	34	31	24	-36	4	22
70°	40	29	8	-37	13	-13
65°	18	1	9	-29	5	-40
60°	2	-2	16	-30	-1	-41
55°	-4	4	10	11	0	-40
50°	-8	2	5	35	-13	-14
45°	29	-3	-12	42	13	6
40°	19	-17	-12	33	26	5
35°	21	-28	-6	31	38	5
30°	36	-40	-6	30	16	-2
25°	43	-41	0	7	-7	12
20°	28	-23	-2	-8	-32	34
15°	24	-7	-4	-5	-32	26
10°	14	12	0	-2	-14	10
5°N	10	21	4	10	6	-2
0°	14	12	9	18	14	-12

5. COMPARISONS

5.1 Comparison with the stratospheric circulation in the southern hemisphere

Obasi (1963) determined the mean meridional circulations in the southern hemisphere at several tropospheric levels and at the stratospheric levels 100 and 50 mb. These $[\overline{v}]$ values derived from actual wind observations, are shown in Table 4 and can be compared with our winter, summer and yearly averages in Table 5. However, there is a slight difference in the periods considered;

in our study:	winter N.H.	= Oct 1957 - March 1958
	summer N.H.	= July-Sept 1957 + April-June 1958
and in Obasi's study:	winter S.H.	= April - Sept 1958
	summer S.H.	= Jan - March 1958 + Oct - Dec 1958

In the study of the southern hemisphere, mainly 00Z data are used; in the present study 00Z and 12Z data are utilized. Some systematic diurnal effects, which are minimized in our final results due to averaging of the 00Z and 12Z observations, could still be important in the southern hemispheric data.

Cross sections of the mean meridional circulations over the entire globe for winter, summer and year are constructed in figures 6 and 7. The thickness of the 100mb-30mb layer is of course very much exaggerated in the diagram compared with the distance to the center of the earth.

The following is of interest to note:

1. The mean north-south circulation in the southern hemisphere is practically the mirror image of the corresponding circulation in the northern hemisphere.

2. The circulation in the northern hemisphere is approximately twice as strong during the winter (N. H.) as during the summer (N. H.). However, in the southern hemisphere the situation is different; the circulations are stronger during the summer (S. H.).

3. Mean southward motion is observed across the equator during the winter N. H. (-10 cm/sec); there appears to be no mean exchange between both hemispheres during the summer N. H.

4. Equatorward flow dominates at 50 and 30 mb, except in the area within 30° of both poles. Near the poles a strong poleward motion is found at all levels.

The yearly averages of the mean meridional circulations for the northern hemisphere are presented in Table 5b. These yearly values were computed as the average of the seasonal $[\overline{v}]$, which are given in Table 1 (i. e. July - Sept 1957, Oct - Dec 1957, Jan - March 1958, April - June 1958 and the times 00Z and 12Z). Under the assumptions that the 00Z and 12Z data are independent and that the data form a Gaussian distribution, the 95 % confidence limits of the yearly $[\overline{v}]$

were computed from the variation of the seasonal $[\overline{V}]$. The 95% confidence limits are given by the expression $\pm \frac{2\,\sigma(x)}{\sqrt{N}}$. Since we did not assume any seasonal trend in our data, the estimate of the standard error will be on the pessimistic side.

Table 4. Summer, winter and yearly averages of the mean meridional velocity $[\overline{v}]$ at 100 and 50 mb for the southern hemisphere*. Units are cm/sec. Mainly 00Z observations are used.

	Winter 1958 (April-Sept 1958)		Summer 1958 (Jan-March 1958 + Oct-Dec. 1958)		Year 1958 (Jan-Dec 1958)	
	100 mb	50 mb	100 mb	50 mb	100 mb	50 mb
$80^{\circ}S$	-39	2	-104	-106	-71	-52
$75^{\circ}S$	-30	-8	-115	-90	-72	-49
$70^{\circ}S$	-9	-24	-63	-71	-36	-47
$65^{\circ}S$	-8	0	17	-33	4	-16
$60^{\circ}S$	-7	21	29	-16	11	2
$55^{\circ}S$	-2	35	32	-2	15	16
$50^{\circ}S$	19	41	47	16	33	28
$45^{\circ}S$	46	53	62	24	54	38
$40^{\circ}S$	62	74	86	41	74	57
$35^{\circ}S$	62	91	114	46	88	68
$30^{\circ}S$	44	34	92	49	68	41
$25^{\circ}S$	21	6	61	46	41	26
$20^{\circ}S$	11	0	29	31	20	15
$15^{\circ}S$	-9	0	14	9	2	4
$10^{\circ}S$	-9	-6	11	-8	1	-7
$5^{\circ}S$	0	-10	-7	-20	-3	-15
0°	7	-7	-12	-36	-2	-21

*From G. O. P. Obasi, 1963: Atmospheric momentum and energy calculations for the southern hemisphere during the IGY.

Table 5a. Summer and winter averages of the mean meridional velocity $[\overline{V}]$ at 100, 50 and 30 mb for the northern hemisphere. Units are cm/sec. 00Z and 12Z are averaged.

	Summer (July-Sept'57 + April-June'58)			Winter (Oct'57-March'58)		
	100 mb	50 mb	30 mb	100 mb	50 mb	30 mb
80°N	35	37	4	70	42	24
75°N	24	42	-2	90	46	32
70°N	19	28	-13	110	51	17
65°N	18	-1	-19	102	26	9
60°N	2	-18	-22	53	-7	-11
55°N	-8	-10	-15	7	-21	-42
50°N	-12	-1	-10	0	-21	-54
45°N	-21	-15	-14	-19	-23	-44
40°N	-36	-24	-18	-56	-25	-37
35°N	-38	-30	-10	-60	-21	-24
30°N	-27	-16	-6	-45	-9	-6
25°N	-36	-7	-9	-4	-3	1
20°N	-32	1	-11	16	-5	-2
15°N	-14	0	-3	12	-13	-11
10°N	-2	2	3	10	-21	-17
5°N	0	0	-1	13	-17	-22
0°	-4	5	-2	14	-4	-14

Table 5b. Yearly average of $[\overline{V}]$ and twice the standard error of the mean for the seasonal $[\overline{V}]$ from its yearly average at 100, 50 and 30 mb for the northern hemisphere. Units are cm/sec.

	July 1957 - June 1958		
	100 mb	50 mb	30 mb
80°N	53 ± 21	40 ± 24	14 ± 20
75°N	57 ± 29	44 ± 19	15 ± 28
70°N	64 ± 38	40 ± 15	2 ± 30
65°N	60 ± 33	12 ± 12	-5 ± 21
60°N	28 ± 26	-13 ± 10	-16 ± 15
55°N	0 ± 11	-16 ± 9	-28 ± 24
50°N	-6 ± 14	-11 ± 15	-32 ± 19
45°N	-20 ± 18	-19 ± 11	-29 ± 13
40°N	-46 ± 18	-25 ± 13	-27 ± 17
35°N	-49 ± 21	-25 ± 17	-17 ± 18
30°N	-36 ± 23	-13 ± 16	-6 ± 13
25°N	-20 ± 27	-5 ± 15	-4 ± 11
20°N	-8 ± 24	-2 ± 13	-6 ± 15
15°N	-1 ± 21	-6 ± 16	-7 ± 15
10°N	4 ± 17	-9 ± 17	-7 ± 14
5°N	6 ± 12	-8 ± 13	-12 ± 11
0°	5 ± 13	0 ± 7	-8 ± 8

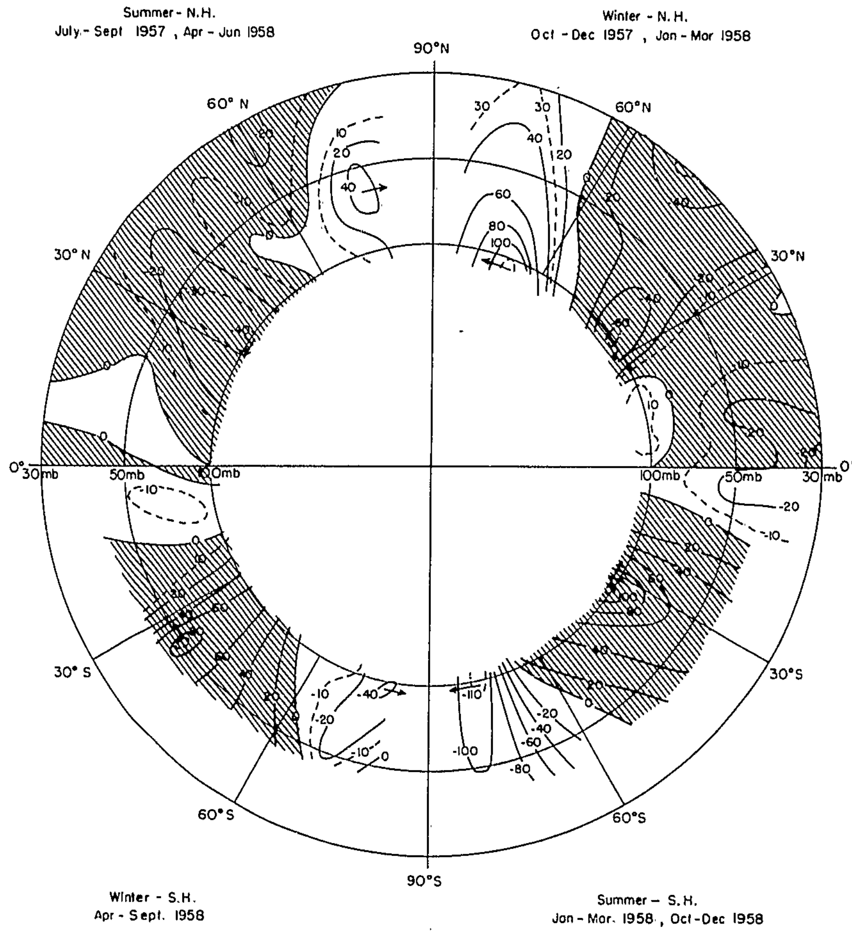

Figure 6. Cross section of $[\overline{V}]$ through the stratosphere (100 - 30 mb) of both hemispheres for winter and summer 1957/1958. The regions of equatorward motion are shaded. Units are cm/sec.

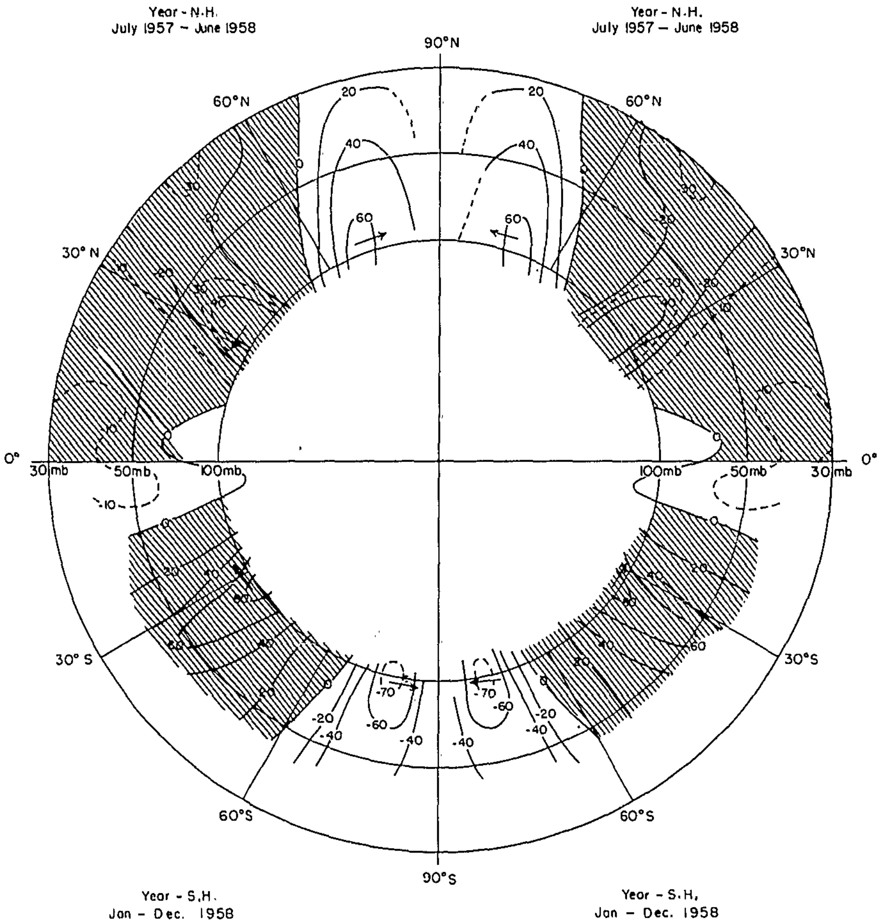

Figure 7. Cross section of $[\overline{V}]$ through the stratosphere (100 - 30 mb) of both hemispheres for the year 1957/1958. The regions of equatorward motion are shaded. Units are cm/sec.

5.2 Comparison with a rotating tank experiment

Starr and Long (1953) have described an interesting dishpan experiment. By a careful choice of the characteristic parameters (depth of the fluid, temperature gradient between rim and center of the tank and the rate of rotation), they tried to simulate the actual conditions in the atmosphere, They measured velocities at the top of the rotating fluid and computed eddy momentum fluxes and mean horizontal velocities ($\overline{[u]}$ and $\overline{[v]}$). There is a difference in the order of averaging in the rotating tank experiment and in our study. Starr and Long computed "daily" $[v]$ and averaged the $[v]$ values for 108 "days" (experiments) in order to determine $\overline{[v]}$. We first took in our own study averages with respect to time and next with respect to longitude: $[\bar{v}]$. In the case of perfect data both methods will give the same result: $\overline{[v]} = [\bar{v}]$.

A difficulty is that it is not known with what level in the atmosphere one can identify the free surface of the fluid. In spite of this uncertainty, a comparison is drawn in Table 6 between the $[\bar{v}]$ at 30 mb in the stratosphere and the $\overline{[v]}$ measured at the top of the rotating fluid.

Although it is obvious that there are many points of difference, it is interesting to note that in almost every experiment rimward

velocities are found at the top of the fluid, similar in character to the equatorward circulations observed at 30 mb in the stratosphere.

Table 6. Comparison of the mean horizontal velocities*, measured in a rotating dishpan experiment at the top of the fluid, and in the stratosphere at 30 mb. The radius of the dishpan was 15 cm.

ROTATING TANK			STRATOSPHERE (30 mb)		
AVERAGE OF 108 PHOTOGRAPHS			AVERAGE OF 1 YEAR OF DATA		
RADIUS	$\overline{[v]}$	$\overline{[u]}$	LATITUDE	$[\overline{v}]$	$[\overline{u}]$
2.50 cm	-.0029 cm/sec	.0705 cm/sec	75^{o}N	+14 cm/sec	303 cm/sec
5.00	-.0043	.1413	60^{o}	-16	554
7.50	-.0051	.2449	45^{o}	-29	784
10.00	-.0099	.4119	30^{o}	- 6	-506
12.50	+.0004	.7290	15^{o}	- 7	-810
13.75	-.0061	.6375	10^{o}	- 7	-565

*Mr. R. E. Dickinson kindly permitted me to use the mean zonal velocities $[\overline{u}]$ which he obtained for the year July 1957 - June 1958.

6. CONCLUSIONS

Due to the improved observations and the expanded station network during the IGY, it is probable that the measured horizontal meridional velocities are representative of the IGY stratospheric circulations. However, more years of data in the troposphere as well as in the stratosphere are needed to verify our results and to draw a definite conclusion about the structure of meridional circulations in the atmosphere. An extension of the present study to wind observations at the four standard hours each day could possibly yield valuable information concerning the semidiurnal tides.

In the yearly average (July 1957 - June 1958) a three cell pattern is observed at 100 mb with poleward motion north of 55°N (maximum at 70°N, + 60 cm/s), equatorward motion from 15°N to 55°N (maximum at 35°N, -50 cm/s) and weak poleward motion from 0° to 15°N (maximum +6 cm/s). At 30 mb the velocities are all to the south (about - 20 cm/s), except in the area poleward of 70°N (maximum of + 15 cm/s). The north-south stratospheric circulation in the summer is generally weaker than during the winter.

It is a well known fact that ozone is accumulated at high latitudes during the early spring (e. g. Craig 1950). Since our data give a mean motion towards the south at 50 and 30 mb, an ozone build up by a mean meridional circulation is not consistent with these results.

Kuo (1956) studied the mean meridional circulation produced by sources of heat and momentum, using Buch's wind data (1954) for the year 1950. Kuo calculates a maximum value of 1 m/s for the meridional velocities. This compares well with the maximum velocity of + 60 cm/s observed in our yearly average. Dickinson (1963) and Teweles (1963) computed $[\overline{V}]$ from momentum convergence and divergence, <u>neglecting small scale friction.</u> They compute velocities of the order of at most 20 cm/s (i. e. a factor 2 or 3 smaller than our data indicate). The direction of the meridional cells which are measured by Dickinson and Teweles agrees, however, with our findings. The similarity in strength and pattern of the circulation of the southern hemisphere (Obasi 1963) forms other evidence in support of the $[\overline{V}]$ pattern presented in this paper.

ACKNOWLEDGEMENTS

I should like to thank Professor Victor P. Starr for his encouragement and advice throughout the course of this study.

I am indebted to Dr. A. A. Barnes, Jr., for writing the computer program and to Dr. G. O. P. Obasi and Mr. R. E. Dickinson for valuable discussions; to Professor R. E. Newell and Mr. J. A. Young for reading the manuscript and their helpful suggestions. Thanks are due to Miss Eileen Gannon for typing the manuscript and to Mrs. Barbara Goodwin and the computing staff of the M. I. T. Planetary Circulations Project for performing the calculations upon which this work is based, and also to Miss Isabel Kole for drafting the maps and figures presented in this report. I am indebted to the "Netherlands Organization for Advancement of Pure Research (Z. W. O.)" for financial support in granting a "NATO Science Fellowship".

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TRANSFER THROUGH THE TROPOPAUSE AND WITHIN THE STRATOSPHERE*

by

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ABSTRACT

The evidence for tropospheric-stratospheric mass exchange is reviewed. A large fraction of the exchange appears to occur in the vicinity of the middle latitude tropopause discontinuity and the concomitant jet stream. Calculations of the flux of radioactive substances in the troposphere support this view. Computations of the horizontal flux of ozone in the lower stratosphere are presented for the IGY period divided into three-month seasons. Indications are that the large-scale quasi-horizontal transient eddies can transport ozone polewards in sufficient quantities to account for the spring build-up of ozone. Such large-scale mixing, as opposed to mean meridional motions, also allows explanation of the distribution of radioactive tungsten in the stratosphere. Transports appear to be polewards and downwards between latitudes 20° and 60°. Independent meteorological evidence, in the form of isentropic cross-sections and values of the covariance between the meridional and vertical components of the wind, support the mixing scheme. The observed counter-gradient eddy flux of heat can be explained from the model. Distributions of certain other trace substances are not at variance with the model. From a preliminary examination of the angular momentum transport processes up to 60 km it appears that transient eddies are also important to these levels but there is not yet sufficient global coverage to evaluate the contributions due to mean motions and standing eddies.

* This paper was submitted to the Royal Meteorological Society in December 1961

1. Introduction

Meteorologists presently claim an understanding of the behaviour of the atmosphere below the tropopause; but comparable claims cannot yet be made for the stratosphere, the mesosphere and the thermosphere. As with most physical systems our comprehension is limited by our ability to observe events properly. In the past few years this ability has increased by leaps and bounds. Reliable data to about 30 km (10 mb) are now gathered on a routine basis from the radiosonde network; with special attention balloons may be used to altitudes of 35 or 40 km. (Conover, Lowenthal and Taylor 1960; Hopper and Laby 1960). The Meteorological Rocket Network over North America provides synoptic wind data, and some temperature data, in four periods of each year at altitudes to 65 km (0.1 mb) (Webb, Hubert, Miller and Spurling 1961). A tantalizing glimpse of the wind and temperature structure to 85 km ($\sim$ 0.004 mb) has been provided by the rocket grenade experiments (Stroud, Nordberg, Bandeen, Bartman and Titus 1960; Groves 1960; Nordberg and Stroud 1961; Teweles 1961) which have shown variability in the mesosphere corresponding to that found in tropospheric weather. Between 80 and 100 km meteor trail drift observations (Elford 1959; Greenhow and Neufeld 1955, 1956), taken over several years, have provided the wind velocities for one station in each hemisphere, including the tidal components whose origin is still not completely resolved. Just above this layer ionospheric discontinuities have been tracked and interpreted in terms of winds (Mitra, 1949). Sodium vapour trails extending to altitudes of 200 km have also been tracked to

obtain wind velocity in a few cases (Manring and Bedinger 1960). Certain rocket flights (Horowitz and LaGow 1957, 1958; Horowitz, LaGow and Giuliani 1959) have enabled temperatures to be calculated, with various assumptions, up to these levels. At higher altitudes, to about 1000 km, our knowledge is restricted to that of density (King-Hele 1961) and composition (Istomin 1959). Again under certain assumptions temperatures can be calculated.

The atmospheric circulation patterns derived from the whole gamut of techniques used below 100 km have been presented by Murgatroyd (1957) and Batten (1961); wind data are more abundant than temperature data in the section above 30 km. The patterns above the tropopause suggest to the meteorologist very similar questions to those that were raised many years ago concerning the circulation of the troposphere. One may enquire, for example, about the source of the kinetic energy represented by these circulations. How much of the kinetic energy is advected across the lower and upper boundaries, either as kinetic or potential or internal energy, and how much is generated in situ from the effects of solar and infrared radiation? This particular topic was the subject of a recent paper by Professor Starr (1959) in which was emphasized the importance of a study of the energy flux across levels such as the tropopause. Another striking fact is the reversal of the wind direction that takes place between summer and winter in the region below about 70 km. Where are the sources and sinks of relative angular momentum for this region? Does the summer easterly regime correspond to some systematic removal of angular momentum

from the region into the troposphere? Or are there some torques, not yet understood, that are operating in situ? One may also enquire about the character of the mixing processes in the region. If a foreign trace substance is introduced into the atmosphere at say 60 km what happens to it? Does it rapidly mix throughout the region 50-80 km, so that in the course of a month or so it is uniformly distributed? Or is it confined to a thin layer in the vertical and limited in latitudinal extent? In contrast to the opinions prevalent some years ago it seems that the atmosphere is reasonably well mixed up to about 100 km (Meadows and Townsend 1960) but the time scale associated with the mixing is not yet known.

Answers to the first two sets of questions can be obtained by application of the principles of conservation of energy and momentum. The third set indirectly can be attacked with the principle of the conservation of mass. One of the first steps should be the construction of a budget to keep track of the energy and angular momentum involved in the region from the tropopause to 100 km. When one constructs from the circulation patterns cross-sections of the kinetic energy density and angular momentum density of the region it immediately becomes obvious that from physical considerations the kinetic energy and momentum advected across the upper boundary make a negligible contribution to the total budget. Such statements cannot at present be made about the higher layers, for example 900-1000 km. It is equally clear that relatively small changes near the lower boundary can give rise to enormous changes in the upper half of the layer considered - if indeed there is any relationship between

events. It therefore seems logical that we build from our knowledge of the troposphere to gain some understanding of the vertical fluxes of energy angular momentum and mass at the tropopause, then investigate in detail the next layer above, say from the tropopause to 30 km, with the ultimate aim that we shall evaluate the vertical fluxes at 30 km as soon as events in the layer are understood and as soon as observations allow.

During the past three years the Planetary Circulations Project at the Massachusetts Institute of Technology, directed by Professor V.P. Starr, has been evaluating the horizontal and vertical fluxes of energy and momentum in the layer from 100 to 10 mb. Data collected during the International Geophysical Year have been used. As is well known similar computations have been made over the past twelve years for the atmosphere between the surface and 100 mb. A picture of the workings of the stratospheric region is emerging that is not only of considerable interest in its own right but which will provide an indispensable springboard for the study of the layers above - say 30-50 km, 50-80 km and 80-100 km. Some of the findings will be quoted below.

There are several approaches to the study of transport processes within, and into, the stratosphere. If direct observations of temperature, pressure and wind velocity are available, one can either attempt a comprehensive study of events over a short period by isentropic trajectory analysis or some similar technique, or one can collect the observations together over a long period of time and study the average values both of

the elements themselves and of derived quantities such as the transport of momentum and energy. The latter, the climatological approach, is the one adopted by the Planetary Circulations Project. Yet another approach is available if one has observations of some trace substance as a function of latitude, longitude, height and time. If the substance can be treated as conservative in its passage from its sources to its sinks then changes in concentration and gradients of concentration can be interpreted in terms of atmospheric transport processes. This technique is essentially an application of the principle of mass conservation.

Until quite recently much of the literature concerning the general circulation of the lower stratosphere has contained conclusions based on observations of the distribution of trace substances. Craig (1948) for example interpreted ozone observations as implying a mean meridional motion from equatorial to polar regions in the lower stratosphere. Brewer (1949) from water vapour and ozone data suggested that there is a direct mean meridional cell involved with rising motion through the tropical tropopause, northward motion (in the northern hemisphere) between low and middle or high latitudes and sinking motion to the north. Similar types of mean meridional circulations have been suggested by Goldie (1950) from meteorological observations, Dobson (1956) from ozone and water vapour data, Stewart, Osmond, Crooks and Fisher (1957), Machta (1957), Dyer and Yeo (1960) Libby and Palmer (1960) all from observations of fission products, and Burton and Stewart (1960) from observations of natural radioactive substances. Murgatroyd and Singleton (1961) have deduced the existence of such a circulation in meridional planes from calculations based on

the radiation budget alone with consideration of possible eddy heat transport omitted. Brewer, Machta, and Murgatroyd and Singleton have pointed out that one of the major drawbacks of these schemes is that the angular momentum budget of the atmosphere is not balanced by the circulations postulated, at least in certain regions.

There is of course a parallel between the suggestions of a mean meridional circulation in the stratosphere and the similar suggestions made many years ago of a direct meridional circulation in the troposphere. Such a direct circulation was supposed to transfer heat energy from the source regions at low latitudes to the sink regions near the poles. Later it turned out that the heat transfer mechanism was somewhat different. Certainly the differential heating produces available potential energy but instead of this being realised as kinetic energy of a mean meridional motion it appears that under the combined influence of the differential heating and rotation of the earth the potential energy is converted into available potential energy of the large scale quasi-horizontal eddies, thence into kinetic energy of the eddies and thence into kinetic energy of the mean zonal flow. It is these large-scale eddies which produce the heat flux poleward and not the mean motion. Indeed these eddies in the process also transport relative angular momentum northward and for a complete angular momentum balance an indirect mean meridional cell has been postulated (Eady 1950) and observed (Starr 1954). In view of the experience with the troposphere it is thus not necessarily logical to suggest that the differential heating in the stratosphere and mesosphere produces a direct mean meridional circulation. Murgatroyd and

Singleton in fact carefully pointed out that eddy transports would ultimately have to be included in their calculations.

Several authors have proposed that the distribution of tracers can best be explained by large scale eddy-mixing processes with a flow down the concentration gradient for any particular tracer. Reed (1953) felt that vertical eddy-mixing was important but that horizontal eddy mixing should also be considered. Martin (1956) was one of the first to investigate the latter suggestion with regard to ozone; his work was based on computations of the horizontal ozone flux using actual ozone and wind data. Godson (1960) and Ramanathan and Kulkarni (1960) also pointed out that baroclinic waves may govern the ozone flux. In a thesis written in 1959 the author (Newell 1960a; 1960b) pointed out that Martin's work could possibly explain the observed distributions of fission product radioactivity and simultaneously account for the stratospheric countergradient flux of heat reported by White (1954). Evidence in favour of the idea came when observations of tungsten 185, collected by U-2 aircraft at altitudes up to 70,000 ft, were released (Feely and Spar 1960). There is now at hand evidence from extensive meteorological data that the eddy mixing interpretation gives a stratospheric model which is not at variance with considerations of angular momentum and energy balance. Some of these data have been discussed elsewhere (Newell 1961). The main purpose of the present paper is to summarize the work on tracers that has led to the current picture of stratospheric-tropospheric exchange and to the current ideas concerning transport processes within the stratosphere. Calculations of the ozone

flux during the International Geophysical Year will be presented in detail as the main observational basis for the model and an effort will be made to fit the observations of trace substances into a picture that is also consistent with the extensive meteorological observations processed by the members of the Planetary Circulations Project.

2. Tropospheric-Stratospheric Interchange from Tracer Studies

The exchange of air between the trosposphere and stratosphere is a topic which is still not thoroughly understood but which has received considerable attention of late in attempts to account for observations of trace substances in the two regions. One of the first studies of the exchange which used methods of synoptic meteorology was that of Reed and Sanders (1953); in evaluating the mechanisms and motions that occurred in the formation of a baroclinic frontal zone they found evidence that stratospheric air was entrained into the zone. Sawyer (1955) in a study of detailed aircraft observations of a frontal zone noticed that a tongue of very dry air was often present in association with the zone and on one occasion was able to trace it backwards in time along a quasi-horizontal path into the stratosphere. In similar vein Ramanathan (1956) suggested that ozone may escape into the troposphere via the quasi-horizontal circulations associated with the jet stream.

Intense interest in the topic was aroused when it was found that many months after nuclear weapons tests had ceased considerable concentrations of radioactivity were still observed in the troposphere in spite

of the fact that the debris had, supposedly, a mean life in the troposphere of only 30 days (the removal being principally by precipitation). The debris originated from these parts of the original nuclear clouds that had penetrated into the stratosphere; such high yield explosions occurred either at high or low latitudes. Machta (1957) suggested that the material entered the troposphere in the vicinity of the middle latitude tropopause discontinuity. Some of the data on the radioactivity of the air will be examined below to see if there is any support for the hypothesis.

One of the best sets of observations of the fission product radioactive substances in the lower troposphere is that collected by the United States Naval Research Laboratory. There are 20 stations along 80°W. At each station air is blown through a circular piece of asbestos-based filter paper, 8 inches in diameter, at a flow rate of about 30 cubic feet per minute. The paper has essentially 100% retention for 0.3 μ particles and 90% retention for particles as small as 0.02 μ. Filters are changed at 0800 local time daily and forwarded to Washington where they are ashed at 650°C and then all counted for gross β -activity with one and the same end-window Geiger-Muller Tube. Each day the count of a standard sample and the background count is obtained. Most stations commenced daily sampling at the beginning of the International Geophysical Year period and continued until November 1959 at which point the low levels of activity made necessary a change to three-day sampling periods. Dr. L.B. Lockhart of the Naval Research Laboratory has kindly supplied the author

with this most valuable geophysical record. Monthly mean meridional profiles have been constructed and are shown in Fig. 1 (see also Lockhart and Patterson 1960; Lockhart, Patterson, Saunders and Black 1960). Tests ceased on November 4, 1958 and apart from the French tests in February, April and December 1960 and April 1961, which are mostly reflected directly in the monthly data, most of the tropospheric radioactivity during 1959, 1960 and 1961 had come from the stratosphere. As already mentioned the mean residence time of radioactive particles in the troposphere is about 30 days (Stewart, Crooks and Fisher 1956). It is very difficult to make a detailed interpretation of these profiles as they are essentially the end product of stratospheric transport processes, stratospheric-tropospheric exchange, tropospheric transport and tropospheric removal processes. All of the factors may vary with latitude and season. Nevertheless, three general points are outstanding; there are maxima in middle latitudes of both hemispheres; the maxima move north and south with the sun; the concentrations are highest in the spring.

The first point cannot be explained by the variation of rainfall with latitude. In the tropics there is an inverse relationship between the radioactivity of the air and the rainfall (Lockhart, Baus and Blifford 1959) but it cannot be extended to middle latitudes for in the periods examined in detail there was more rainfall at the stations with high activity (25-40^{o}N) than in the region to the north with much lower activity. The strontium 90 content of soil samples taken from the entire globe show a similar middle latitude maximum; even when they are collected

from a line of equal rainfall amount (Alexander 1959). In an effort to elucidate the second point, profiles of mean monthly zonal component of the 200 mb wind at stations along 80°W in the northern hemisphere were compared with the radioactivity profiles and it was noted that the respective maxima move north and south in phase with the radioactivity maxima displaced to the south by 5-10°. The displacement is in the direction that would be expected from physical considerations as the potential temperature surfaces from the lower middle latitude stratosphere that penetrate the tropical troposphere or baroclinic zone slope downwards towards the equator. It is not intended to assert that the transport is purely meridional. The secondary maximum in the northern hemisphere at Thule (74°N) that appears after tests at high latitudes by the U.S.S.R. represents either direct stratospheric-tropospheric exchange at high latitudes or a longer wash-out time.

There is clearly much more information that can be gleaned from the concentration values and no doubt will be gained when the injection into, and removal from, the troposphere are more thoroughly understood. For the present the very simple assumption will be made that the daily concentration values along 80°W in the northern hemisphere are not directly related to the rainfall. The actual tropospheric wind data can then be used to estimate the flux of fission products in the lower troposphere. Surface concentrations were combined with wind velocities from 1.0 or 1.5 km and calculations were made of the north-south eddy flux of fission products by means of the technique that will be outlined

later in the discussion of ozone. The average monthly fission product concentration at 4000 feet has a correlation coefficient with the concentration at the surface of +0.85 (calculated from data tabulated by Pierson, Crooks and Fisher 1960) and thus it seemed reasonable to take the transports as representative of the lower troposphere. From the flux values at adjacent stations along 80^{o}W, with the assumption that the zonal flux is divergence free, the divergence of radioactivity in a given volume of the lower troposphere was calculated. The equation of continuity was then used to estimate the vertical flux into or out of the volume. If convergence was indicated it could only come about by descent of material from above whereas divergence could be interpreted either as removal by precipitation or settling on to the earth's surface or as upward transport. Four two-month periods free of tests in the winters of 1958 and 1959 were chosen for the calculations. The results showed a divergence of the meridional flux from the region of 30-35^{o}N which suggests that downward transport occurred into that region. There was also an indication of a southward transport from high latitudes which may have corresponded to direct stratospheric-tropospheric exchange. The results have been discussed in detail elsewhere (Newell 1960b) and have since been extended to include the continent of North America. Both zonal and meridional fluxes have been calculated for the entire year 1959. Again in the region of 80^{o}W there are indications of downward transport in middle latitudes with the region of downward flux moving northward in summer. A complete description of this more complex procedure

will be published in a separate paper. The main object here is to point out that not only the daily values themselves but also the tropospheric transport values derived from these values can be explained quite well with the assumption that the majority of the radioactivity enters the troposphere in the region of the baroclinic zones and tropopause discontinuities. The possibility of a secondary direct source at high latitudes from tests by the U.S.S.R. is quite reasonable on meteorological grounds as the lapse rate in the lower stratosphere in winter is negative as in the upper troposphere. Presumably small-scale vertical eddy mixing can proceed more easily in these conditions than when there is an inversion at the tropopause (see for example the work of Ball 1960).

There have recently been several detailed studies of the structure of the isentropic surfaces in the vicinity of frontal zones over North America. These studies (see Reed and Danielsen, 1959; Danielsen 1959; Staley 1960) show several good examples where isentropic surfaces pass from troposphere to stratosphere. Consideration of constant potential vorticity trajectories on these surfaces shows the physical possibility of transfer in both directions. The laminar structure revealed suggests that transfer is accomplished quasi-horizontally rather than by direct vertical circulations.

Having established that, according to the meteorological analysis, air can pass between the two regions in the vicinity of the jet stream and baroclinic zone and having shown that such passage provides a reasonable explanation for the high concentrations of fission products observed in

the middle latitude troposphere one must enquire if there is any direct evidence of the transfer of trace substances in the region of interest. Sawyer (1955) showed cases where the dry air in the frontal zone had probably come from the stratosphere. Helliwell's (1960) measurements suggest transfer in the opposite sense as he reports several occasions when the air of the lower stratosphere was relatively rich in water vapour in the vicinity of frontal zones. Brewer (1960), in a discussion of some of the measurements of ozone concentration made with his chemical sonde, has pointed out several ways that ozone may be transferred from the stratosphere into the troposphere including direct transfer downwards through the tropopause and also the schemes presently under discussion. His measurements show evidence of ozone-poor layers in the lower temperate stratosphere which may have come from the troposphere and there are also indications of ozone-rich layers in the troposphere. Ney and Kroening (1961) using the chemiluminescent type of ozone sonde described by Regener (1960) have also detected ozone-rich layers, about 1 km thick, in the upper troposphere which may have entered from the stratosphere. It would seem that the quasi-horizontal motions indicated from the work of Danielson and Staley could adequately account for these findings.

Roach (1961) has recently presented some meridional cross-sections of ozone and frost-point constructed from observations made by the Meteorological Research Flight. The isopleths of ozone concentration show a definite protrusion from the stratosphere into the troposphere in the region of the jet stream, the baroclinic zone and the tropopause gap which are all at the same general latitude. The effect is much more pronounced

in winter. Roach also confirmed Murgatroyd's (1959) finding of a high negative correlation between water vapour and ozone in the vicinity of the jet stream. Ozone-poor water vapour-rich layers in the stratosphere had probably recently been in the troposphere.

Giles (1961) has summarized the data on the concentration of strontium-90 in the vicinity of jet streams obtained in the troposphere and stratosphere from air samples collected by aircraft. Again there is a tendency for the isolines to follow the tropopauses with higher concentrations above these surfaces and there is a protrusion into the troposphere in the vicinity of the jet stream. Paetzold and Piscaler (1961) have reported the protrusion evident from ozone soundings made during and after the International Geophysical Year. Their observations suggest protrusions at both the polar and the sub-tropical jets in winter.

In summarizing the status of the tropospheric-stratospheric exchange problem it can be said that there is now considerable experimental evidence from the observations of trace substances which indicates that much of the mass exchange occurs in the region of the jet stream and tropopause discontinuity. Meteorological evidence in the form of isentropic cross-sectional diagrams is not at variance with this view. Direct vertical exchange at the tropopause or exchange by virtue of the local change of height of the tropopause are factors whose contribution is not yet known. Since many of the isolines of the trace substances tend to parallel the tropopause away from the regions of the jets, it would appear that these factors are smaller than the mass exchange discussed above.

The first two facets of the surface-air radioactivity values have now been discussed at some length. The third aspect was the higher concentrations observed in the spring. Reference to Fig. 1 shows that the phenomenon is best marked in the northern hemisphere in the spring of 1959. In 1960 events were confused by the French tests of February and April but nevertheless a rise in levels appeared in December 1959 and January 1960. In 1961 there was again confusion from the French tests of December 1960 and April 1961 but yields were low and again the rise appeared at stations not influenced by these tests. It is actually possible by using isotope ratio data or age determinations to eliminate the confusion entirely but the pertinent data will not be discussed here. Suffice it to say that a seasonal effect of meteorological origin has been established. One may ask whether the effect is due to increased stratospheric-tropospheric exchange in the winter and spring or to increased stratospheric mixing that leads to higher concentrations in the middle latitude stratosphere or to both causes. The fact that ozone amounts are a maximum in the spring suggests that stratospheric transport is the principal variable. An attempt to examine the stratospheric mixing process will be outlined in the next section.

3. Transport of Ozone within the Stratosphere

The problems relating to the distribution of ozone in the atmosphere have been discussed at length by many authors (see for example Dutsch 1946, 1956; Craig 1950; Normand 1953; Paetzold 1956; Godson 1960; Martin 1956, Martin and Brewer 1959) and as they have had considerable airing in the pages of this journal no attempt will be made to present a comprehensive review. The photochemical theory of ozone predicts maximum amounts in the layer between 20 and 30 km, with total amounts in a vertical column decreasing from the equator polewards and from summer to winter. Observations of the vertical distribution of ozone confirm the presence of the layer but observations of the horizontal distribution show an increase in the total amount of ozone with latitude in winter and higher values in middle and high latitudes in the late winter and spring than in the late summer and autumn. The time necessary to reach 50% of the photochemical equilibrium concentration, after a disturbance in which the ozone content of an air parcel is entirely removed, is 30 minutes, three days, and seven months, for heights of 50, 30 and 20 km respectively for the overhead sun, and 90 minutes, one month and 12 years for the sun at the horizon (Nicolet, 1958). Hence it is obvious that the concentration of ozone in the layers below about 30 km is not influenced directly by the sun. The air in these regions is effectively shielded from the ultraviolet radiation and the ozone can thus be treated, in a sense, as a conservative tracer. Theory and observations can be reconciled if atmospheric circulations are postulated which transport ozone from low

to middle and high latitudes or downwards out of the photochemical equilibrium regions in middle and high latitudes. Such atmospheric circulations have already been referred to in the introduction. In the present section calculations of the horizontal flux of ozone within the stratosphere will be presented and an effort made to assess the role of quasi-horizontal circulations in the ozone budget.

Suppose that the ozone concentration at a particular point at a given time is O ; we can write

$$O = \overline{O} + O'$$

where the bar represents the time mean and the prime the departure from the mean.

Likewise the northward component of the wind V may be written

$$V = \overline{V} + V'$$

The instantaneous northward transport of ozone will be given by

$$OV = \overline{O}\,\overline{V} + O'V' + \overline{O}\,V' + O'\overline{V}$$

and the time averaged transport by

$$\overline{OV} = \overline{O}\,\overline{V} + \overline{O'V'}$$

The equation is an expression of the fact that the northward transport of ozone at a particular point is due to transport by mean meridional motions $\overline{O}\,\overline{V}$ and transport by transient eddy processes $\overline{O'V'}$. For the latter term to be effective in the present problem it is necessary that northward moving parcels of air contain more ozone than southward moving parcels. The eastward transport of ozone could

be written similarly as

$$\overline{OU} = \bar{O}\,\bar{u} + \overline{O'u'}$$

If the discussion is extended from a point to the hemisphere and the northward transport at a particular level is considered, then a third possibility exists for meridional transport over extended time periods, namely, that in the pattern of the so called standing eddy disturbances around the globe, the large scale troughs and ridges, there exists some systematic relationship between the meridional components of the wind and the ozone concentration. For example, if there is more ozone in the region to the east of a trough line and to the west of a ridge line than elsewhere then a northward transport of ozone must exist. Formally such a transport can be expressed as $\left[\bar{O}\,\bar{V}\right] - \left[\bar{O}\right]\left[\bar{V}\right]$ where the square brackets represent average values around latitude circles. These procedures whereby a transport process is resolved into three components, the mean, the transient eddy and the standing eddy transports have been used extensively in the study of planetary circulations in the past ten years (Priestley 1949; Starr 1951; Starr and White 1952; Starr 1954, 1957, 1959). Originally they were applied to such intrinsic properties of the atmosphere as wind velocity and temperature to calculate momentum and heat transports, but in recent years they have been extended to the study of trace substances such as water vapour (Starr and White 1955; Hutchings 1957), ozone (Martin 1956) and radioactive substances (Newell 1960b).

In order to discuss with rigour the transport of ozone over the hemisphere it is necessary to know the concentration of ozone as a function of latitude, longitude, height and time as well as the concomitant wind distribution including both vertical and horizontal components. In practice the only synoptic observations of ozone on a global basis are those of the total amount of ozone measured by the Dobson spectrophotometer. This total amount refers to all the ozone in a vertical column above the station while a myriad of wind velocities are occurring at the same time in the column. It is difficult to see how to compute the ozone flux from these data. Fortunately it has been found that when the total amount of ozone changes much of the change occurs at least in middle latitudes, between 12 and 24 km. Mateer and Godson (1960) find the coefficient of correlation between the total amount of ozone and the ozone in the 12-24 km layer is +0.97 (for a Canadian station). They find that changes of ozone in this layer account for three-quarters of the total change, on a daily or seasonal basis. The amount of ozone at Canadian stations on which the study was based ranged from about 0.29 to 0.50 cm at STP. Ramanathan (1956) found a similar result for the 18-27 km layer at an Indian station (the ozone centre of gravity in the vertical is higher at low latitudes). There, ozone amounts ranged from 0.15 to 0.21 cm. Vertical distributions in each case were found by the umkehr method. The distributions and findings based on them suggest that a first approximation to the transport of ozone in the stratosphere can be obtained from wind velocities representative of the 12-24 km layer together with a measure of the total amount of ozone.

The Planetary Circulations Project of the Massachusetts Institute of Technology is engaged in an extensive study of the general circulation of the stratosphere, based on observations collected during the International Geophysical Year. Computations of the seasonal wind velocities and temperatures and fluxes of angular momentum and energy are performed by machine methods; a similar approach to that already extensively applied at tropospheric levels (see Starr 1954, 1957) is being used. Data from levels of 100, 50, 30 and 10 mb are used; in the standard atmosphere they correspond to heights of 16, 21, 24 and 30 km. Altogether some 220 stations in the northern hemisphere are used. Twenty five stations which were either at, or close to, good wind reporting sites, reported daily values of the total amount of ozone.

The amounts have been incorporated into the machine program and three-month seasonal averages calculated for the six periods of the IGY. In addition to average values $\overline{O}$, $\overline{V}$, $\overline{u}$, $\overline{T}$ and their standard deviations $\sigma(O)$, $\sigma(V)$, $\sigma(u)$, $\sigma(T)$, covariances $\overline{O'V'}$, $\overline{O'u'}$, $\overline{O'T'}$, $\overline{V'T'}$, $\overline{u'V'}$ and correlation coefficients $r(O,V)$, $r(O,u)$, $r(O,T)$, $r(V,T)$, $r(u,V)$ were calculated. The data which involve ozone values will be presented here, although reference will be made to the other results which will be discussed in detail elsewhere by my colleagues. There are too few data to make an analysis of the 30 and 10 mb results. The fluxes reported refer only to levels of 100 and 50 mb (approximately 16 and 21 km respectively). These are most appropriate levels according to the vertical distribution data examined above.

At the lower latitudes it would perhaps be better to examine only the 50 mb results but there is always present some inter-level wind correlation (Charles 1959) so that both levels will be presented. The ozone stations used are listed in Table 1 together with the wind velocity stations in parentheses where wind and ozone were measured at different sites. There are other ozone stations in the Northern Hemisphere but either ozone or wind values were not available here at the time the computations were made (during 1960 and early 1961). As soon as ozone amounts are made available we propose to extend the computations to 1959, 1960 and 1961.

Wind velocities were received for both 0001 GMT and 1200 GMT for all stations except Marcus Island, which relayed data only for 0001 GMT. All the calculations mentioned above were performed separately for these two times. Ozone stations in Canada, Japan, the United States and the U.S.S.R. reported values of the total amount of ozone applicable to several different times each day (on World Meteorological Organization form 0-1); in such cases the ozone value nearest in time to the wind observation was selected for the computations. Completed forms for Moosonee and Edmonton were not received until after the calculations were made and forms for the European stations still have not been received. In these cases one tentative value was available for each day and was used in the calculations with the two wind velocities nearest in time.

A summary of the average ozone amounts, by station and season, principally compiled from the results of the 1200 GMT calculations, appears in Table 2. Ozone amounts for Tromso (69° 40'N 18° 57'E) are also included for reference; this is of course one of the oldest ozone-reporting stations and of great interest but winds were not available to the heights necessary for our transport calculations. The average amounts show the well known seasonal variation although it is not so well marked when three-month averages are considered. Most stations show higher amounts in the January-March 1958 period than the April-June period but the reverse is true for the two low latitude stations and for those in the latitude belt from 45-55°N. There is a general tendency for ozone amounts to increase with latitude except in the October-December periods. The variation of ozone amount with longitude is quite striking. Sapporo, at 43°N 141°E has very much higher values in the first half of 1958 than stations in comparable latitudes at other longitudes. It is difficult to decide, on the basis of just one station, whether the high values are meteorological in origin or are due to instrumental effects. A recent paper by Sekiguchi (1961) favours the former explanation.

The standard deviations, also shown in Table 2, are largest in middle and high latitudes and in the January-March 1958 period. In the July-September periods the variability exhibited is low at all latitudes.

Perusal of the covariance values shows considerable variation from one season to the next and even between the same season in successive years. Variations with longitude are also evident. These variations

are not surprising to the meteorologist, particularly when the relatively small number of observations is taken into account. The principal interest in the present paper is in the general circulation of ozone. We have therefore collected together observations in the same latitude belt to form some estimates of the global flux of ozone. Calculations for both time sets have been combined and thus the number of cases quoted does not always represent the number of independent observations but may be up to a factor of two higher than this quantity. There was an exception to the grouping by latitude belts in the case of Japan. For reasons to be discussed presently it was decided to treat these three stations as a separate group rather than include them in the latitude belts with others. There were not sufficient stations at high latitudes to warrant combination into groups. Covariance values, henceforth referred to as transient eddy fluxes, are shown in Table 3.

The levels 50 mb and 100 mb are both presented for comparison. The resemblance between the results reflects the inter-level wind correlation. At low latitudes it is more pertinent to consider the 50 mb flux as the 100 mb level is often within the troposphere. The 50 mb level fluxes are generally positive in middle latitudes with largest values appearing in the spring. The Japanese stations show a fairly strong negative flux in the same period. Keflavik, unfortunately the only good ozone and wind station north of 60^{o}, too shows a very strong negative flux in the spring. At 50 mb the one low latitude station shows a northward flux throughout the year. When the average seasonal

winds over Japan were studied it was found that there is a strong jet over the area in the spring and autumn; it is possible that the high negative transports actually represent ozone amounts being transported downwards into the troposphere in the vicinity of the jet as discussed in section 2. The very high values of ozone amount in the area may also be due to these circumstances.

There is considerable interest attached to these two findings - that the transient eddy flux is generally northward and that the maximum is in the spring season - for these are precisely the nature of the transports required to reconcile theory and observations as pointed out earlier. But it is necessary to establish that these transports are of sufficient magnitude to be of importance in the global balance of ozone. For this purpose a crude ozone budget of the stratosphere has been constructed.

The contribution of mean motions and standing eddies to the budget are, with the present number of stations, very difficult to assess. One of my colleagues on the Planetary Circulations Project has recently completed a study of the first six months of stratospheric data from the IGY and has constructed charts of the seasonal averages of the meridional wind component which have been used to estimate the average values of the wind round latitude circles. For the six-month period July-December 1957 these values (see Barnes 1961) show a southward motion between 25° and 55°N with a maximum magnitude of about 15 cm/sec, and a northward motion with a maximum of 10 cm/sec between the equator and

$25^{o}N$ at 50 mb. The values are not strictly comparable with the ozone fluxes for they represent a six-month period. Even using all the available data from 220 stations in the northern hemisphere Barnes found that the hemispheric mean meridional motions were so small that they were almost lost in the meteorological "noise" unless two three-month periods were combined. At some later date we shall have available seasonal averages for several years and then this problem will perhaps have been circumvented.

A purely objective appraisal of the contribution to the ozone flux by standing eddies appears in Table 4 as calculated directly from the formula shown. Stations in the 40-$60^{o}N$ latitude belt, widely spaced in longitude, were included. There is an indication of northward transport which again apparently has a maximum in winter and spring. Once more we must stress the crudeness of this approach - ideally we would like to use many more stations - but this seems to be the only step possible at the present time.

To obtain an independent estimate of the seasonal changes of ozone amount, the curves presented by Godson (1960) of the average total ozone amount as a function of latitude and month of the year were used. The ozone amount from the pole to $50^{o}N$ and $40^{o}N$ was integrated on a monthly basis and three monthly changes were averaged to give the average change in ozone content of the region during the season. For comparison with the January-March fluxes, the content change from December to January, January to February and February to March was computed. The budget

calculations for the 50 mb level are shown in Table 5. In order to convert from the fluxes recorded in Tables 3 and 4 to actual ozone transports it was assumed that 35% of the total amount of ozone was actually involved in the layer in which transport was occurring centred on 50 mb. This approximate figure was obtained from the summaries of umkehr observations presented by Ramanathan (1956), Dutsch (1959) and Mateer and Godson (1960). Mr. Barnes' values of the mean meridional motion were used to calculate the flux in the July-September period. The actual transports given by the direct calculation of the standing eddy transport are also included. For reasons which have already been discussed both of these transports should be viewed with caution. Perhaps the most important point to note is that the transient eddy flux, by itself, is of the same order of magnitude as the observed change in ozone content of the polar cap in the January-March period. Thus the large-scale quasi-horizontal transient eddies may be the prime factors in the movement of ozone northwards in the stratosphere.

From ozone concentration measurements taken near the earth's surface Regener (1957) has been able to estimate the vertical flux of ozone in the troposphere; his value is 1.2×10^{11} molecules cm^{-2} sec^{-1}. The measurements were in middle latitudes. Let us suppose that a flux one-half as great is representative of the entire hemisphere; the middle latitude values might be expected to be higher than elsewhere if most of the ozone enters the troposphere in the vicinity of the tropopause gap. Let us further suppose that all the ozone flowing downwards in

the troposphere initially flowed northwards quasi-horizontally across an imaginary vertical boundary along $40^{\circ}N$; in this case the equivalent flux in the stratosphere corresponding to the ozone flux at the ozone sink is $\sim 6 \times 10^{9}$ cm^{-3} sec^{-1} as shown in the Table. It is of the same order of magnitude as the calculated fluxes and provides an independent piece of evidence that these fluxes are important in the general circulation of ozone.

If we combine the results of the ozone budget estimate with those of section 2 of this paper we can hypothesize the following transport path for ozone. Ozone moves northwards from low to middle and high latitudes in the lower stratosphere. There it is subjected to vertical mixing and can reach the isentropes which sometimes pass into the troposphere in the vicinity of the jet stream. Once there, it can be removed from the stratosphere. As mentioned earlier the Japanese stations may reflect this exchange. The largest northward transport of ozone into the middle latitude stratosphere occurs in the period January-March and the well-known maximum in the total amount appears also at this time at the high latitude stations. Afterwards much of this ozone is presumably lost to the troposphere. Junge (1961a) has recently collected values of the tropospheric concentration of ozone and shown that this too exhibits a seasonal variation with the maximum in the tropospheric concentration at the surface being 1-2 months later than the maximum in the total amount of ozone. Junge's finding fits very nicely with the present model. It would thus seem that greater stratospheric mixing in winter is the reason

for the observed spring maxima in both fission products and ozone.

The hypothetical path followed as outlined above does not seem to be subject to any long time delays. After the increased mixing in the stratosphere there soon appears a maximum value in the ozone amounts followed one or two months later by higher values at the surface as the ozone passes into the troposphere. The speed of these eddy processes can be gauged from the length of time necessary to replenish the polar cap. It has already been demonstrated that the increases in the polar cap content can be accounted for on the basis of the calculated eddy flux values. If the ozone in the region from the pole to 50°N from the tropopause to 25 km were completely removed and then replenished by the transient eddy flux it could attain the maximum content (the value for March) in just under four months. Similarly if the maximum polar cap content passed into the troposphere through the tropopause gaps at one-half the rate equivalent to the flux measured by Regener it would take about four months to remove all the ozone.

4. Transport of Radioactive Tungsten within the Stratosphere

Several moderate-yield tests in the equatorial Pacific in the summer of 1958 produced radioactive tungsten 185 (whose half life is 74 days) which was injected into the troposphere and lower stratosphere. Tungsten is not a fission product; it was produced in the bombardment of tungsten 184 by neutrons. It is believed that the sole source was the 1958 summer test series by the United States and the isotope

therefore serves as a unique tracer for the debris from this series. It was first produced on May 13 or 14, 1958 and was detected in surface air by the Naval Research Laboratory stations along 80°W before the end of May (Lockhart, Baus, Patterson and Saunders, 1960). By the end of July the profile along 80°W appeared very similar to that usually observed for gross fission products; there were peaks in both hemispheres, as can be seen in Fig. 2, some 20-30° north and south of the latitude of injection at 11°N. Lockhart et al consider that the transports northward and southward to the maxima occurred in the upper troposphere. It is possible that the transport may have occurred in the lower stratosphere in which case the two maxima represent regions of stratospheric-tropospheric exchange. Stebbins (1959) has pointed out that the first of the radioactive clouds which contained tungsten moved westwards round the globe in the stratosphere to reach eastern United States some 44 days after the explosion during which time the debris spread in latitude by about 45°. The equivalent meridional velocity is about 1.2 metres per second which is large enough to support the argument that the two peaks came from stratospheric debris.

The data of Fig. 2 show an increase in the surface air concentration of tungsten in the spring of 1959. Similar spring maxima had been noted earlier in the concentration of gross fission products and in the rainwater content of radioactivity but interpretation had been difficult because of the pattern of the tests. Martell (1959) considered that spring maxima were principally due to the removal from the polar stratosphere

of debris introduced by Russian tests in the previous autumn. The tungsten measurements demonstrate that debris whose ultimate source is the equatorial stratosphere also enters the troposphere at a greater rate during the spring. Lockhart et al, from considerations of the ratio of strontium 89 to tungsten 185 (Sr^{89} has a half life of 50.5 days) show that debris younger than that from the 1958 summer series also is involved in the 1959 spring maximum. Walton (1960), who has measured the concentration of tungsten 185 and strontium 90 in rainfall, estimates that only 8% of the total strontium 90 in rainfall collected in the spring of 1959 was from the equatorial tests - the remainder being from the October 1958 Soviet tests in the Arctic. Lockhart, Patterson, Saunders and Black (1960) estimate a value of 10% for the same fraction. The principal point to note is that debris from both stratospheric sources arrives at middle latitude and produces a maximum in the spring. A polewards mean meridional motion is difficult to reconcile with these facts.

The various arguments involving fission product radioactivity have so far referred to measurements made at the earth's surface, except the work reported by Stebbins (1959). In 1960 a considerable body of data was released on the concentration of radioactivity in the stratosphere (Stebbins 1960; Feely and Spar 1960). Samples of stratospheric particulates were collected over the Americas to heights of 70,000 ft by Lockheed U-2 aircraft. Two examples of the stratospheric content of tungsten 185 are shown in Figs. 3 and 4. Concentrations represent average values over the two month periods shown. The figures have recently been published

by Stebbins (1961). In an earlier publication on this topic (Newell, 1961) we have reproduced similar diagrams, with permission from Dr. Feely, which were kindly provided to us by Major Stebbins. All tungsten 185 disintegration rates are corrected for radioactive decay back to August 15, 1958. As a background we have used isentropic cross-sections for July 1957 and December 1957 taken from the work of Taylor (1960). The tropopause positions are from Taylor's work. The Planetary Circulations Project has data for the Northern Hemisphere only at present. In the previous superposition isentropes for the stratosphere of the northern hemisphere for longitude 80^{o}W have been used (Newell 1961). Mean isentropes for the hemisphere give a very similar picture. Ultimately our work will provide isentropes for 1958 but for the present paper the temperature cross-sections for 1957 will be compared with the tungsten concentrations for 1958. Both figures suggest that tungsten was transported polewards more or less along the isentropic surfaces with the greatest transport occurring in the winter hemisphere. As Feely and Spar point out indications are that the motion is some type of turbulent mixing rather than an organized mean meridional circulation. The zone of maximum concentration rather than moving northwards with time, as would be expected for a mean meridional motion, actually moved southwards. The two secondary maxima are possibly formed by the removal of tungsten into the troposphere in the region of the tropopause gaps. The secondary maxima have almost identical potential temperatures in both periods. Between the two periods the equivalent potential temperatures of the maxima decreased, in spite of the fact that the southern

hemisphere maximum actually rose in altitude while that in northern hemisphere sank. Because of the complexities involved, in the interpretation it is not possible to argue unequivocally that the decrease is a diabatic cooling. Gravitational settling in conjunction with large scale diabatic vertical motion is another possibility - as also is the chance that coincidence played a part. Table 6 shows the terminal velocities of small spherical particles with a density of 2 gm cm^{-3}. The values were obtained from a graph published by Junge, Chagnon and Manson (1961). The distance fallen by these particles in two months, the time between the tungsten cross-sections, is shown in Table 7. Although the actual distribution of radioactivity among the stratospheric particle size spectrum is not well known at present it appears that radioactive material is present on particles in the size range from 0.01 μ to 1.0 μ. (Stebbins 1960). If it turns out that most of the activity is on the smallest particles in this range then essentially none of the vertical displacement of the tungsten is due to gravitational settling; but as the table illustrates it will be necessary actually to measure the distribution of the radioactivity as a function of particle size before a ruling can be made on the matter. In later periods the maxima depart quite chaotically from the mean isentropes given here (see Stebbins, 1961). It is part of our plan to examine the actual isentropes along longitudes close to the collection sites as soon as the data are processed.

When the tungsten data were first made available in late 1960, it was very gratifying to us to see the general agreement between our

indirect findings based on ozone calculations and the direct measurements in the stratosphere. The fact that eddy transports were predominant in both cases and that transport was polewards and downwards at least to about 60°N, gave reason to believe that a beginning had been made in the formulation of a model of stratospheric transport processes that was consistent with meteorological observations (to be mentioned later) as well as with measurements of trace substances. A more detailed study cannot be made until meteorological data for other seasons has been processed and until the distribution of ozone in three dimensions is at hand. Such steps may take several more years. A few rather generalized comparisons between meteorological evidence, tungsten 185 and ozone observations will be presented below.

Let us suppose that the meridional displacement of the mid-latitude maximum in the northern hemisphere or the displacement of the 5000 dpm/1000 SCF contour between the two profiles gives a representation of the meridional component of the quasi-horizontal eddy speed. For the region 40-60°N a speed of about 21 cm sec^{-1} is obtained. In like fashion the equivalent speed for the ozone flux may be obtained by dividing the calculated flux by the average ozone amount. Such a procedure yields a value of about 17 cm sec^{-1} from the transient eddy fluxes averaged over the two October-December periods in the 45-60°N latitude belts.

Spar, quoted by Stebbins (1960), has used the tungsten profiles and a quasi-Gaussian model of turbulent diffusion to estimate diffusion rates in both vertical and horizontal directions. He finds a value for

the horizontal diffusion coefficient of 10^9 cm^2 sec^{-1}. A value for the same parameter in the troposphere is 10^8 - 10^{10} cm^2 sec^{-1} obtained by Grimminger (1941) from a study of isentropic charts. The difference between the two regions lies not so much in their horizontal motions, for it is well known that strong winds with a certain degree of variability exist in the stratosphere, as in their vertical structure. Measures of the variability will be quoted later.

In principle an equivalent eddy diffusion coefficient could be calculated from the ozone flux calculations in conjunction with the observed ozone distributions. Some of these distributions have been presented by Ramanathan and Kulkarni (1960) from umkehr measurements and by Brewer (1960) and Paitzold and Piscaler (1961) from measurements with ozone sondes. But there are not really enough soundings available yet to make quantitative estimates of diffusion from such profiles.

Spar's estimates of the vertical diffusion coefficient, which are somewhat more difficult to make as the source had a finite height, are 10^3 cm^2 sec^{-1} for the tropical stratosphere and about 4×10^4 cm^2 sec^{-1} for middle latitudes. The difference is in the direction that would be expected on meteorological grounds as the temperature increases with height in the tropical stratosphere whereas in middle and high latitudes it increases by a much smaller amount and in the winter season at high latitudes it decreases with height.

The difference in the vertical diffusion coefficients in the two regions of the stratosphere may be an important factor in the physical

explanation of how the ozone budget can be balanced by the quasi-horizontal motions discussed. Ozone can diffuse down the concentration gradient and therefore northwards and downwards as long as appropriate eddy shuffling occurs apparently following the general pattern of the isentropic surfaces. The process can only continue as long as ozone is removed from the region to which it is transported. The large vertical eddy diffusion allows ozone to be transferred vertically, also down the gradient, so that a particular column at middle latitudes can actually build up more ozone than a similar column at low latitudes where vertical diffusion is so much smaller. In this way ozone is apparently transferred northwards against the gradient but the gradient now is envisaged as that due to total amounts of ozone; actually quasi-horizontal eddies are transporting ozone down the concentration gradient in a particular isentropic layer. It is possible in this fashion to remove one of the major objections to the quasi-horizontal transfer approach which has hitherto been that ozone could not be transferred against the gradient of total ozone. Such thinking has led to the conception that ozone must come into the column from above. Of course in the simple picture outlined here no attempt has been made to specify the actual processes which produce the vertical eddy diffusion - they may just be the vertical components of what have been referred to as large-scale quasi-horizontal eddies. Undoubtedly a particular isentrope has a considerable vertical displacement both with time at a given position and with latitude and longitude at a given time.

But events cannot be completely isentropic, otherwise debris would remain in the stratosphere indefinitely. There is a general tendency for the debris to reach lower isentropes where it has access to the troposphere. Another point evident from the tungsten cross-sections is that inter-hemispheric mixing can occur fairly easily in the lower stratosphere. This is not the case in the lower troposphere as is evident from the radioactivity measurements shown in Fig. 1, although Lockhart et al suggest that tropospheric inter-hemispheric exchange can occur at certain times. Consideration of the whole set of tungsten cross-sections shows that more poleward transfer occurs in the winter hemisphere.

Although the tungsten data provide good evidence that mean meridional motions are not the major contributor to the poleward flux of material it is by no means ruled out that small meridional motions cannot exist at certain times. The latitude of the maximum concentration of tungsten as a function of time taken from the cross-sections published by Stebbins (1961) is shown in Table 8. During the first winter there was apparently a southward drift with a corresponding velocity of 12 cm sec^{-1} followed by a northward drift. But because the maximum was for the majority of the time just above the sampling altitudes it is very difficult to interpret this as a mean meridional motion. Another point is that the tungsten data were collected between longitudes of 70^{o} and $130^{o}W$ and do not therefore represent a global mean. It is quite probable that large scale standing eddies exist in the tungsten concentration just as they do for ozone or wind velocity and in such a case a phase change of the standing eddy pattern could be interpreted as a mean meridional motion.

5. Meteorological Evidence of Meridional Transport

In the January-March period it has been shown that both transient and standing eddies transport ozone northwards. Northward moving parcels must therefore contain more ozone than southward moving parcels. The implication from the ozone calculations taken alone was that the transport is horizontal. An equally fair interpretation would be that the transports are quasi-horizontal and that northward moving parcels are descending so that they tend to be removed from the layer where photochemical equilibrium prevails. The tungsten data too have revealed polewards transports that are also downwards and the configuration of the isentropes leads to the suggestion that these motions are, to a certain extent, isentropic.

One might ask if there is any meteorological evidence of such an effect, namely that northward-moving parcels are sinking and southward moving parcels are rising in the middle latitude lower stratosphere. White (1954) reported a counter-gradient eddy flux of heat at 200 mb and 100 mb at latitudes 31°, 42.5°, 55° and 70°N. The exact mechanism whereby this flux came about was unknown at the time but in light of the ozone and tungsten data a possible interpretation is that northward moving parcels are sinking and therefore warming adiabatically and southward moving parcels are rising and cooling. In Table 9 are shown average values of the covariance between the 50 mb meridional velocity and temperature at the ozone stations used. Time, seasonal and longitudinal averages are performed in exactly the same way as for the ozone flux data in Table 3.

At 50 mb in the latitude range from 30° to 60°N the covariance is positive and this is just what would be expected in view of the configuration of the isentropes if the motions are largely adiabatic. The same configuration leads one to suspect that the particularly strong negative value at the Icelandic station in the spring and winter of 1958 corresponds to northward moving parcels ascending. It will be recalled from Table 3 that in the same seasons the station exhibited a southward flux of ozone by transient eddies. As there is little information on the actual distribution of ozone north of this station it is not possible to ascribe the flux values to an increase of ozone concentration with latitude or altitude.

Further evidence has come from some recent work here by Loisel and Molla (1961). They have evaluated the covariance between northward and vertical components of the motion in the northern hemisphere using the vertical velocities computed by Jensen (1960) by the adiabatic method together with the horizontal components of the wind. Their results for the transient eddies are quoted, with permission, in Table 10. In the troposphere the covariance values are positive indicating that in general northward moving parcels are rising and southward moving parcels are sinking. In the 50-100 mb layer, however, the sign reverses and it appears that northward moving parcels are sinking south of 50°N and rising to the north of this latitude.

Individual trajectories of air parcels over Europe drawn by Martin and Brewer (1959) in their study of ozone changes showed, contrary

to their expectations, that at the end points of northward trajectories local changes of total ozone and temperature were both positive. Their findings could be explained by the mechanisms discussed above in which northward moving ozone-rich parcels are subsiding and warming adiabatically. Although no attempt has been made here to apply the climatological findings to individual situations it seems that the relationships between temperature and ozone amount discussed by Godson (1960) for such situations are not at variance with the present interpretation. The covariance between 50 mb temperature and ozone amount is shown in Table 11 in the same form as the previous covariance summaries. Positive values are predominant, this circumstance being presumably due to the descent and adiabatic heating of ozone-rich parcels. Ultimately it will be desirable to examine the diabatic factors involved in the temperature changes. It should be noted from Table 11 that the largest covariance values occur in the spring. Calculations performed for Tromso (not shown) for 100 mb also showed the same sign as the Icelandic station with large values in the spring. It should be noted that Meetham's (1938) early work, in which a positive correlation between total amount of ozone and the potential temperature at 18 km was found, can also be interpreted by these same mechanisms.

Ozone and tungsten measurements have shown transports from which one could infer greater quasi-horizontal shuffling in the stratosphere in winter than in summer. A meteorological measure of the shuffling might be given by the variance of the meridional component of the wind.

Dr. Takio Murakami of the Planetary Circulations Project has recently completed a summary of the hemispherical averages of these variances as a function of latitude for the first two three-month periods of the IGY and he was kind enough to allow me to peruse the values which he will be publishing shortly. At 50 mb, during July-September 1957, the variances were about 3 m sec^{-1} and showed little dependence on latitude with a maximum close to 4 m sec^{-1} near 25°N. In the October-December 1957 period, variances were again about 3 m sec^{-1} near the equator but increased to about 7 m sec^{-1} near 65°N with values of about 5 m sec^{-1} near 80°N. The main points to note are the increase of variance in the winter season and the increase from the equator polewards in this season. Both points substantiate the conclusions from the ozone and tungsten data. Such large values make it very difficult to detect a mean meridional motion of only 10 cm sec^{-1} or so such as was discussed in relation to the ozone budget. Mr. Barnes discussed this point in his paper in which the mean meridional motions for the six-month period July-December 1957 were presented.

6. Zonal Flux of Stratospheric Ozone

The results of the calculations concerning the zonal flux of ozone by transient eddies appear in Table 12 for the 50 mb level. While little is presently known about the balance requirements and circulation of ozone in the zonal direction apart from the fact that there are changes in ozone amount from one longitude to another as shown in Table 2, it is possible

that the mean zonal motion and its standing eddies control the zonal flux. The flux ascribed to the transient eddies may be a reflection of the pattern of angular momentum transport insofar as the covariance $\overline{O'u'}$ will be predominantly positive in regions where the covariances $\overline{O'v'}$ and $\overline{u'v'}$ are large and positive. For example the covariance $\overline{O'u'}$ in the October-December 1957 period is positive in middle latitudes as also is the covariance $\overline{O'v'}$. The hemispheric averages show that $\overline{u'v'}$ too is positive in this season, that is relative angular momentum is transported northwards by the transient eddies. The final interpretation of the zonal eddy flux values must therefore await completion of our studies now underway concerning the angular momentum budget of the stratosphere.

7. Application of Findings to Other Trace Substances

(a) Strontium 90.

The distribution of strontium 90, whose half life is about 28 years, shows middle latitude maxima in the soil, surface air and rainwater (Alexander 1959, Lockhart et al 1960, Stewart et al 1958). No attempt will be made to give a comprehensive bibliography of the results. In the stratosphere air samples to 70,000 ft have been collected by the Lockheed U-2 aircraft (Stebbins, 1960, 1961) and the Atomic Energy Commission has used balloons to raise sampling equipment (the Ashcan program) to 100,000 ft (Holland 1959). Recently the United States Weather Bureau has also conducted

sampling missions in the troposphere and lower stratosphere (Giles 1961). The pattern in the stratosphere, somewhat chaotic during the period of intensive weapons testing in 1958, has gradually evolved since that time as the relative contribution of the various groups of tests has altered. Figs. 5 and 6 show the stratospheric distribution during the first six months of 1959 and 1960. In 1959 when the Ashcan balloon program enabled the isolines to be drawn to 100,000 ft, albeit from a somewhat inadequate sample size as far as meteorologists are concerned, there is a broad zone of high concentration stretching from low to high latitudes in the northern hemisphere. Undoubtedly this is a result of the winter eddy mixing of debris from the high and low latitude tests of 1958. The maximum concentrations are at a higher elevation than those of tungsten 185 because the latter was injected by tests of only moderate yield whereas much of the strontium was injected by high-yield tests whose clouds penetrated higher into the stratosphere. The effect of the greater stability and consequently smaller mixing in the lower tropical stratosphere can be clearly seen. In 1960 (Fig. 6) the pattern had changed and there were maxima at middle and high latitudes and at the high altitudes sampled by the U-2 aircraft. Feely and Spar (1960b) and Stebbins (1961) have quoted some work by Kalkstein which shows that the pattern of rhodium 102 was similar to these strontium patterns. The majority of the rhodium, another unique radioactive tracer (half life 210 days) that is not a fission product, was injected by two high yield weapons fired from rockets above 100,000 ft in the equatorial mesosphere in late summer of 1958. The clouds from one of these tests

apparently reached a height of 1,000,000 ft. While interpretation of this tracer is made difficult by its complex characteristics the general pattern of the results is thought to be reliable. The rhodium from the high level explosion first appeared in the stratosphere in the summer of 1959 and by the summer of 1960 there were almost equal amounts in both hemispheres. The interpretation given by Stebbins to both the strontium 90 and rhodium 102 1960 patterns is that they are due to the entry into the sampling network at high latitudes and high altitudes of the debris from the rocket shots. The debris then is supposed to mix equatorwards down the concentration gradient in much the same manner as debris from the equatorial tests mixed polewards. The presence of the region of relatively low mixing in the equatorial stratosphere is again apparent from figure 6. It is presently too difficult to estimate how much of the strontium came from the rocket shot and how much is residual debris from the tests of 1958 and earlier. Further indirect evidence that the debris at high latitudes and altitudes originates from the rocket shots is provided by the age estimates derived from cerium 144 to strontium 90 ratios (cerium 144 has a half life of 285 days). These show that debris in the northern hemisphere is younger over the pole and at high latitudes than over the equator; of course another possible interpretation is that the debris came from high altitude Russian tests in the October 1958 series. The meteorological interpretation of the 1960 strontium and rhodium patterns is that the debris from the explosion in the mesosphere mixed laterally at high altitudes possibly in the mesosphere before mixing vertically down to the sampling regions the vertical mixing being an accompaniment of the winter polar vortices. It is important to

notice here that a model based upon observations of a single element could be quite misleading. In the case of strontium 90 a natural interpretation of Figs. 5 and 6 would be that a mean meridional circulation of the Dobson-Brewer type was in operation. Yet the tungsten 185, ozone and rhodium 102 tracers suggest that this is not the case below 30 mb. It is not at present possible to present a concrete case for either type of circulation at higher levels.

(b) Strontium 89.

The principal use of strontium 89 in tracer studies has been its application to provide a measure of the age of samples. Its half life is 50.5 days and its physical similarity to strontium 90 makes it unlikely that atmospheric events differentiate between the two. It has therefore been the case that the ratio of strontium 89 to strontium 90, in air, water or soil, has been considered as an age parameter. Feely and Spar (1960b) presented values of the ratio in stratospheric air for the 1958-59 period. Values in the equatorial stratosphere, between 60,000 and 70,000 ft decreased with a half time of 51 days between January and September 1959 and when the rate of decrease was extrapolated backwards to a value considered appropriate for the production ratio the date obtained was July 1958 which was the time at which tests were carried out at low latitudes. After September 1958 the ratio no longer decreased. Feely and Spar consider that this may have been due to the difficulty of dealing with very low concentration of strontium 89 or to an influx of younger debris presumably from the October 1958 Russian tests. Samples from the southern hemisphere

stratosphere gave lower values of the ratio between September 1958 and early 1959 indicating the presence of older debris. The ratio increased to reach approximately the same values as the equatorial region by the middle of the year. The increase was gradual which is in accordance with the view that large scale eddy mixing was bringing debris from the equatorial stratosphere into the region.

(c) Carbon 14.

In the stratosphere an excess of radioactive carbon 14, whose half life is 5760 years, over the natural background produced by cosmic rays, has been introduced by the high-yield weapons tests. Hagemann, Gray, Machta and Turkevich (1959) have reported measurements of the concentration of carbon 14 in samples to heights of 100,000 ft collected with the aid of high-altitude balloons. They have constructed a meridional cross-section illustrating the distribution in July 1955. The concentration lines in the stratosphere extend from the equatorial regions downwards and polewards and concentrations increase with altitude. At this time the majority of the bomb carbon had come from tests conducted at low latitudes. A mechanism which could bring about the observed distribution from the single source is the quasi-horizontal large-scale eddy mixing referred to above; in fact this was mentioned by the authors but at the time they favoured an interpretation in terms of the Brewer-Dobson direct mean meridional circulation.

(d) Beryllium-7.

Cosmic rays are thought to be the sole source of beryllium-7 in the atmosphere. The half life of beryllium-7 is 53 days. The theoretical concentrations have been compared with those observed by the U-2 sampling network by Stebbins (1961). Both vertical and horizontal gradients are less steep than those predicted and amounts are generally lower except in the equatorial stratosphere. Again eddy mixing provides a good explanation for the difference whereas, as Stebbins points out, a mean meridional motion model would predict higher concentrations in the polar regions rather than the observed lower amounts.

(e) Radon and its daughter products.

A tracer offering the possibility of examining exchange from the troposphere to the stratosphere is radon 222, a gas with half life of 3.8 days exuded from rocks in the earth's crust, and its daughter products, Radium A, B and C, all particulates with very short half lives, Radium D (or lead 210 as it more usually termed) with a half life of 19.4 years, Radium E with half life of 5 days and Radium F with half life of 138 days which decays to stable lead 206. Measurements of Radium D and F in air in the troposphere and lower stratosphere have been reported by Burton and Stewart (1960). They find that specific concentrations increase with height in the troposphere and increase more sharply with height just above the tropopause. Burton and Stewart interpret the higher values in the lower stratosphere as being due to

transport of radon and its daughter products from the equatorial stratosphere northwards by mean meridional motions of the Dobson-Brewer type. The radon rich air is supposed to enter the stratosphere by vertical motions in the equatorial regions. It is possible to account for the observations equally well with the assumptions that the radon and its daughters enter the stratosphere by quasi-horizontal mixing in the vicinity of the jet stream and tropopause gap and then are transported northwards by eddy mixing as discussed earlier. Under various sets of assumptions Burton and Stewart estimate the circulation time from equatorial to middle latitudes to be between 177 and 212 days. If we suppose that the meridional eddy speeds derived from the ozone and tungsten data are applicable to the radon and its daughters and suppose that about 30° of latitude has to be traversed northwards of the tropopause gap associated with the subtropical jet then we arrive at a transit time of between 173 and 197 days. It would seem that meridional eddy mixing provides an equally valid explanation of the observations. Telegadas and List (1961) have reported some lead-210 concentrations over North America in samples collected in the spring which show much lower concentrations than those over England and show maximum values at 20,000 and 30,000 ft with values at 50,000 ft some 10 times smaller than those over England at that level. In view of this situation and several other complicating factors, such as the possible production of lead 210 in tests by the neutron bombardment of bismuth 209 and lead 208, it is best to postpone further speculation about the role of the atmosphere until more data are forthcoming. One

additional complication that is not usually mentioned in discussions concerning the daughter products of radon is that some become positively charged by recoil on decay and as has been well known for many years (see Rutherford, 1904) they can be collected on a negatively charged wire. The author was fortunate to witness some experiments on the top of Mount Withington, New Mexico by Professor M. H. Wilkening and Dr. A.W. Kawano in which they sampled radon and its daughter products separately at the same site. When thunderstorms were close their electric fields removed all the charged daughter products from the air; the fact that the radon content remained constant demonstrated that the effect was electrical in origin rather than being due directly to air motion.

(f) Water vapour.

While ozone measurements provided the seeds for the Brewer-Dobson circulation model (Dobson, Harrison and Lawrence 1929) the strongest impetus came from the early measurements of the water vapour content of the lower stratosphere. Brewer (1949) reported very dry air just above the troposphere over England with frost points close to -80°C which correspond to mixing ratios of about 2×10^{-3} gm kgm^{-1}. Brewer suggested that the air had been dried by passage through the equatorial troposphere region where temperatures are about -80°C and he theorized that the air then moved northwards in a mean meridional motion and subsided over middle and high latitudes. More recent measurements by the British Meteorological Research Flight (Murgatroyd, Goldsmith and Hollings 1955; Helliwell, Mackenzie and Kerley 1957; Helliwell 1960) have confirmed the presence of

these low frost points. Roach (1961) reports a frost point of -80°C above the polar troposphere in summer, and Roach (1961) and Kerley (1961) report similar low frost points in the upper equatorial troposphere but the latter region is somewhat isolated from the lower troposphere according to their cloud data and has a different lapse rate. These facts are difficult to reconcile with the idea that air in the tropical regions slowly rises through the tropical tropopause. On the basis of present knowledge it would probably be an equally fair interpretation to claim that the atmospheric circulation somehow acts to bring about very low concentrations of water vapour near the equatorial tropopause and that as a consequence very low temperatures appear because there is little water vapour to absorb the outgoing long wave radiation.

Soon after Brewer's report three high altitude balloon flights over North America showed much higher values of the mixing ratio and a tendency for the ratio to increase with altitude above about 15 km to values close to 100 x 10^{-3} gm kgm^{-1} in two out of three cases (Barrett, Herndon and Carter 1950). Such high values were initially viewed with suspicion based mainly on the thought that water vapour may have been taken up by the balloon and apparatus and caused contamination of the high level readings. But recently Mastenbrook and Dinger (1960) reported a similar increase in mixing ratio with height and a value of about 80 x 10^{-3} gm kgm^{-1} at 30 km. At the 100 mb level Mastenbrook and Dinger's data were comparable to the British data. Barclay, Elliot, Goldsmith and Jelley (1960) using a cooled vapour trap,

measured the humidity directly at 27 km and found a value of 37×10^{-3} gm kgm^{-1}. Houghton and Seeley (1960) presented some spectroscopic evidence that is not inconsistent with such values at and above the 27 km level. Murcray, Murcray and Williams (1961) from infrared absorption measurements made with a spectroscope carried aloft on a balloon also found an increase of mixing ratio above 17 km and for the path length above 30 km the absorption corresponded to mixing ratios of about 100×10^{-3} gm kgm^{-1} (if the mixing ratio is supposed independent of height above that level). These authors emphasize the patchiness of the moist layers.

The Japan Meteorological Agency (1961) has reported measurements from a series of special dew point sondes, several of which have reached 10 mb, launched during the IGY and 1959-60. Considerable variability has been revealed; sometimes an increase of dew point does occur at the highest levels with mixing ratios exceeding 100×10^{-3} gm kgm^{-1}, but on other occasions a decrease of mixing ratio with height is noted. Independent evidence of the existence of considerable water vapour above the tropopause is provided by the mother-of-pearl and noctilucent clouds. Mother-of-pearl clouds occur at northern latitudes in winter at levels between 23 and 29 km (Stormer, 1948). Present evidence suggests that they are composed of water even though temperatures are apparently about -75 to -80°C when they are reported. Saturation vapour densities over water are not normally tabulated at such low temperatures, as it is the general opinion that the water will be in the form of ice. Extrapo-

lation from the values calculated for temperatures down to $-50^{\circ}C$ and quoted in the Smithsonian Meteorological Tables (List 1951) gives a vapour density of about 0.004 gm m^{-3} at $-75^{\circ}C$ which corresponds to a mixing ratio at 25 km of about 100×10^{-3} gm kgm^{-1}. The value is the same order of magnitude as the mixing ratios already quoted for these levels. The noctilucent clouds are usually interpreted as either ice or dust (Ludlam 1957). They are seen at high latitudes in summer at levels of about 80 km. The rocket grenade experiments (Nordberg and Stroud 1961) indicate that at just this time and in this region occur the lowest temperatures observed anywhere in the upper atmosphere. Temperatures of $170^{\circ}K$ have been reported. The saturation vapour density over ice at $-103^{\circ}C$ is about 0.000010 gm m^{-3} (List 1951) and the density at 80 km is about 2×10^{-5} kgm m^{-3}, hence a mixing ratio of 500×10^{-3} gm kgm^{-1} is obtained. Although this is higher than the reports at 30 km by a factor of five it is by no means unreasonable. In fact some of the Japanese results approach the value. Furthermore, in the presence of respectable vertical motion such as apparently exists with the clouds (Paton 1954) and which is to be expected in the mesosphere, it is not necessary to have such high humidity values to get cloud formation, as is well known for the case of the troposphere.

Thus there are several lines of evidence that point to the existence of higher water vapour concentrations above 25 km than are found in the lower stratosphere and one of the most intriguing present-day questions about the upper atmosphere concerns the origin of this water vapour.

De Turville (1961) claims that sufficient hydrogen is received at the earth from space to account for all the water present in the oceans. He feels that water may be formed in the higher atmosphere essentially by the oxidation of incoming protons. But although it may be possible to argue that there is a down-gradient flux of water vapour between 30 km and the tropopause it was noted in section 2 that in the vicinity of the tropopause gap water vapour seemed to be entering, rather than leaving, the stratosphere. It is difficult to explain the high values on the basis of large-scale eddy diffusion working in the opposite direction to the ozone flux as the present observations would suggest that such a flux would be counter-gradient. More extensive coverage of the globe geographically and in altitude is required before one can satisfactorily argue this point. Other possibilities are that the water vapour has been introduced to the higher levels by volcanic eruptions and nuclear explosions. Or it may be that we have not yet come to grips with the problem; perhaps the water vapour is bound to the stratospheric particulates such as those studied by Junge, Chagnon and Manson (1961). Junge (1961b) has reported some preliminary evidence that some of the particles are wet. Diffusion may then proceed down the gradient of particulates yet in the opposite direction to the gradient of water vapour concentration. Finally it should be emphasized that there is not a thorough understanding of the radiative effects of the water vapour. It is not meteorologically impossible for the layers or clouds of water vapour to be introduced into the stratosphere near the tropopause gap and to rise diabatically

at certain times and in certain regions by the absorption of long wave radiation. Indeed it is already clear that the radiation budgets constructed by Murgatroyd and Goody (1958) with the assumption of low water vapour concentrations in the stratosphere will have to be revised to take account of the higher concentration of water vapour now thought to be present.

8. Concluding Remarks

At the opening of the essay it was stressed that the logical approach to the study of the high atmosphere is to examine first of all the vertical fluxes of energy, angular momentum and mass at levels such as the tropopause and then to study transports within the layer immediately above, say to 30 km, and so forth. The greater part of the discussion has been concerned with the stratospheric transport and stratospheric-tropospheric exchange of mass. It has been established that a large fraction of the exchange occurs by what are essentially quasi-horizontal exchange processes in the vicinity of the baroclinic zones, the jet streams and the tropopause gaps. Within the lower stratosphere the ozone and tungsten data provide good evidence that large scale quasi-horizontal eddy processes accomplish most of the mixing although the effects of mean meridional motions cannot be ignored entirely. Such processes satisfactorily account for the general distribution of other trace substances with the notable exception of water vapour. There is considerable meteorological evidence which also favours the model, the counter-gradient

heat transport data and the covariance between the meridional and vertical components of the wind being two good examples.

No attempt has been made here to make use of the two other diagnostic tools, namely the principles of conservation of energy and angular momentum. My colleague Mr. Barnes will shortly present a comprehensive study of those terms in the energy balance of the stratosphere that can be determined from meteorological observations and it would seem that there is nothing in the present work that conflicts with his findings. It will be recalled that a preliminary study of a selected situation by White and Nolan (1960) brought out the importance of eddy processes in the energy budget and demonstrated that they acted to convert kinetic to potential energy in the lower stratosphere. Other members of the Planetary Circulations Project are presently giving attention to the various components of the angular momentum budget of the stratosphere. Their work too has not so far produced any evidence of conflict with the circulation processes deduced from the distributions of trace substances, in fact the situation is quite to the contrary. It will be recalled that in the older pictures of the general circulation of the stratosphere a polewards mean motion imposed certain difficulties when the angular momentum balance was considered. The northwards motion actually produced a northwards transport of momentum that was too large on the basis of the observed westerly winds. In the present view there is a southwards mean motion in middle latitudes, rather like that in the upper troposphere. Preliminary evidence shows that in the October-December 1957 period the quasi-horizontal eddies

transport angular momentum northwards south of 60°N and southwards north of this latitude. Thus the eddies act in the right direction to bring about the formation of a polar jet. Although the detailed budget has not yet been drawn up it is clear that the eddy transport coupled with the southward transport by the mean motion does not violate angular momentum considerations.

At the outset it was pointed out that higher layers could be considered in turn. While there are no vertical velocities available for 30 km some wind data have been collected by the Meteorological Rocket Network over North America (Webb, Hubert, Miller and Spurling 1961) at heights to 60 km. Preliminary wind data have been supplied to us very soon after the firings and have been used so far principally in conjunction with class room instruction. While it is appreciated that no firm conclusions can be based on these preliminary data it is of some interest to treat them rather as the stratospheric data have been treated, bearing in mind that it will be many years before good climatological values are available for these levels. The mean zonal and meridional wind velocities for the winter seasons 1959-60 and 1960-61 are shown in Table 13. All data have been assigned either to winter or summer in the summaries. N represents the number of observations. Units are all in knots. It can be seen that there are indications of a mean meridional motion polewards at middle latitudes at about 50 km. A similar motion with smaller magnitude is evident in the summer summary. But caution should be exercised in the interpretation, for this

may just be a reflection of standing eddy pattern such as is observed at lower altitudes. Thus the existence of mean meridional motions at these levels will remain in doubt until the network is extended to Europe and Asia. The covariance of the meridional component $\sigma(v)$ is also shown for cases where 10 or more values were available. The meridional eddy shuffling apparently increases with altitude although part of this variance may be of instrumental origin. Table 14 illustrates the transport of angular momentum calculated from the observations. The important point to note here is that even at 50 km the transport by transient eddies is not negligible compared with the total transport. It is not possible with so few observations to make a start yet on the construction of a budget to 60 km but this will be done in a few more years if observations are continued and extended. This concludes our general argument that eddy mixing processes are important above the tropopause as well as below. The fact that the molecular weight of air remains constant with altitude up to about 100 km is good evidence that mixing occurs up to this level; within the next few years we should reach an understanding as to the amount of the mixing that can be ascribed to large scale eddy motions such as are in evidence at lower levels.

ACKNOWLEDGEMENTS

I am very grateful to Professor V.P. Starr for valuable discussions on the matters dealt with herein. I have also had the benefit of access to the data being processed by colleagues of the Planetary Circulations Project and have enjoyed discussions concerning these data with Dr. T. Murakami and Mr. A.A. Barnes Jr. Mrs. Barbara Goodwin and Mrs. Dorothy Berry have been responsible for the calculations.

The work has been supported jointly by the Atomic Energy Commission under Contract No. AT (30-1) 2241 and by the Geophysics Research Directorate, Air Force Cambridge Research Center, under Contract No. AF 19(604) 5223.

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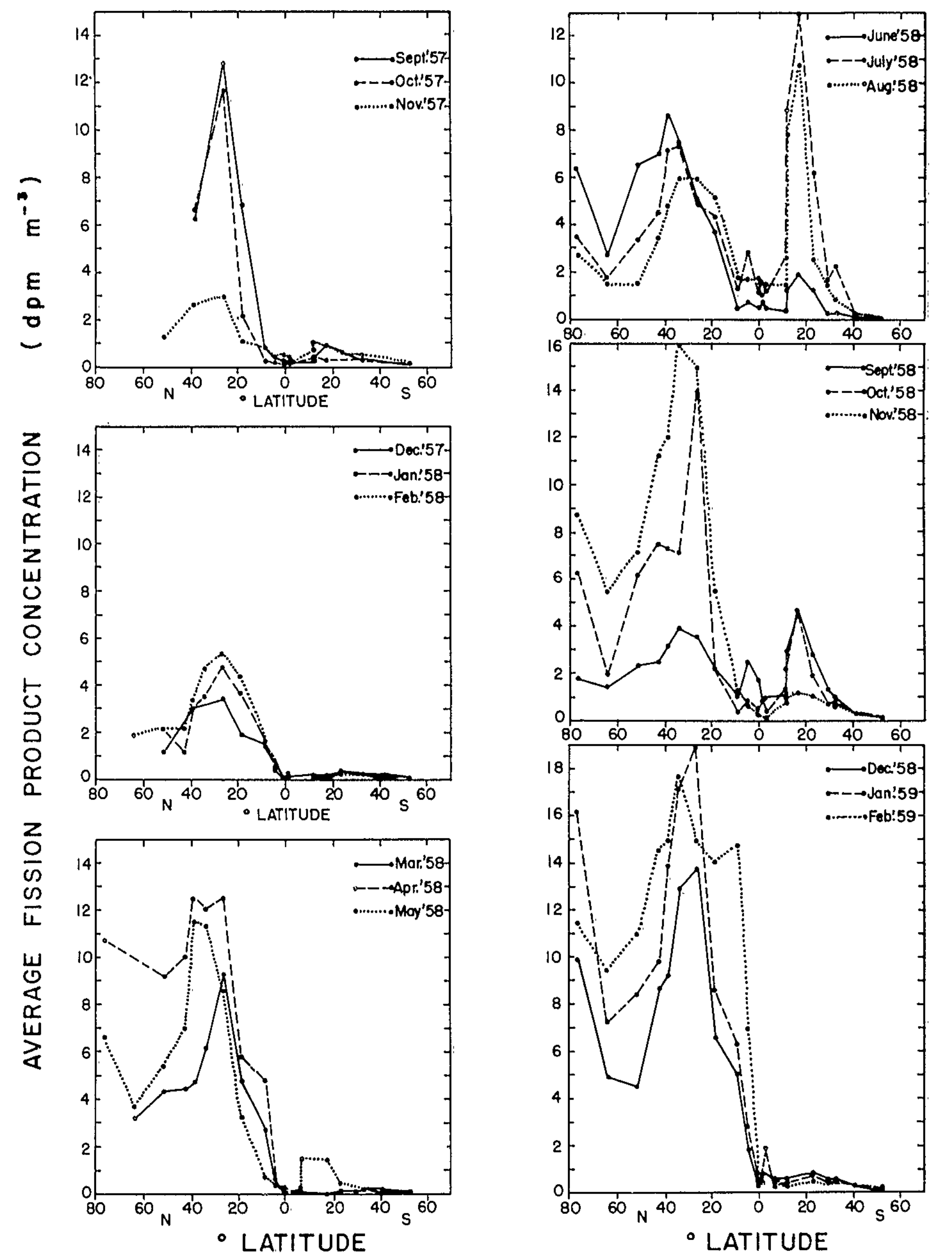
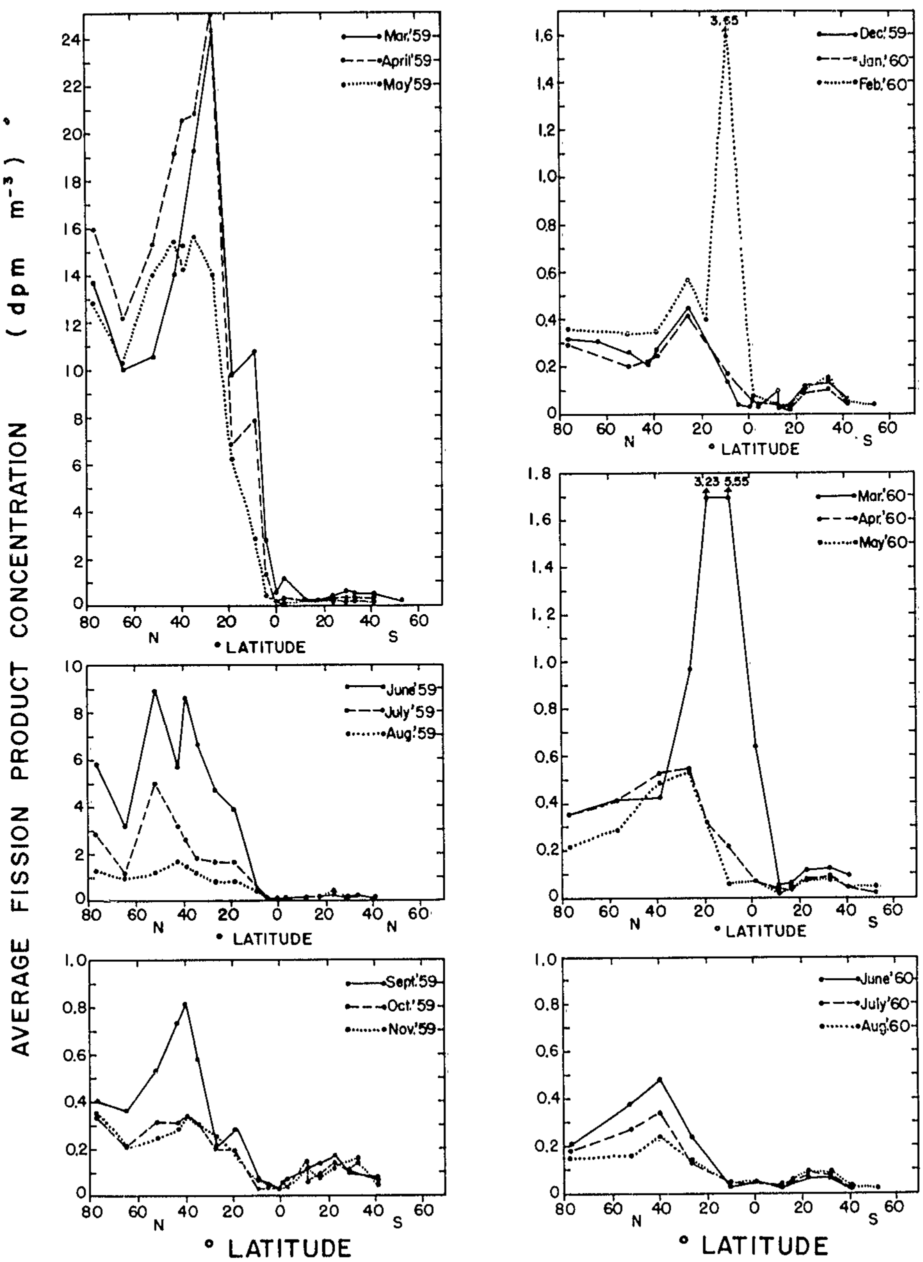
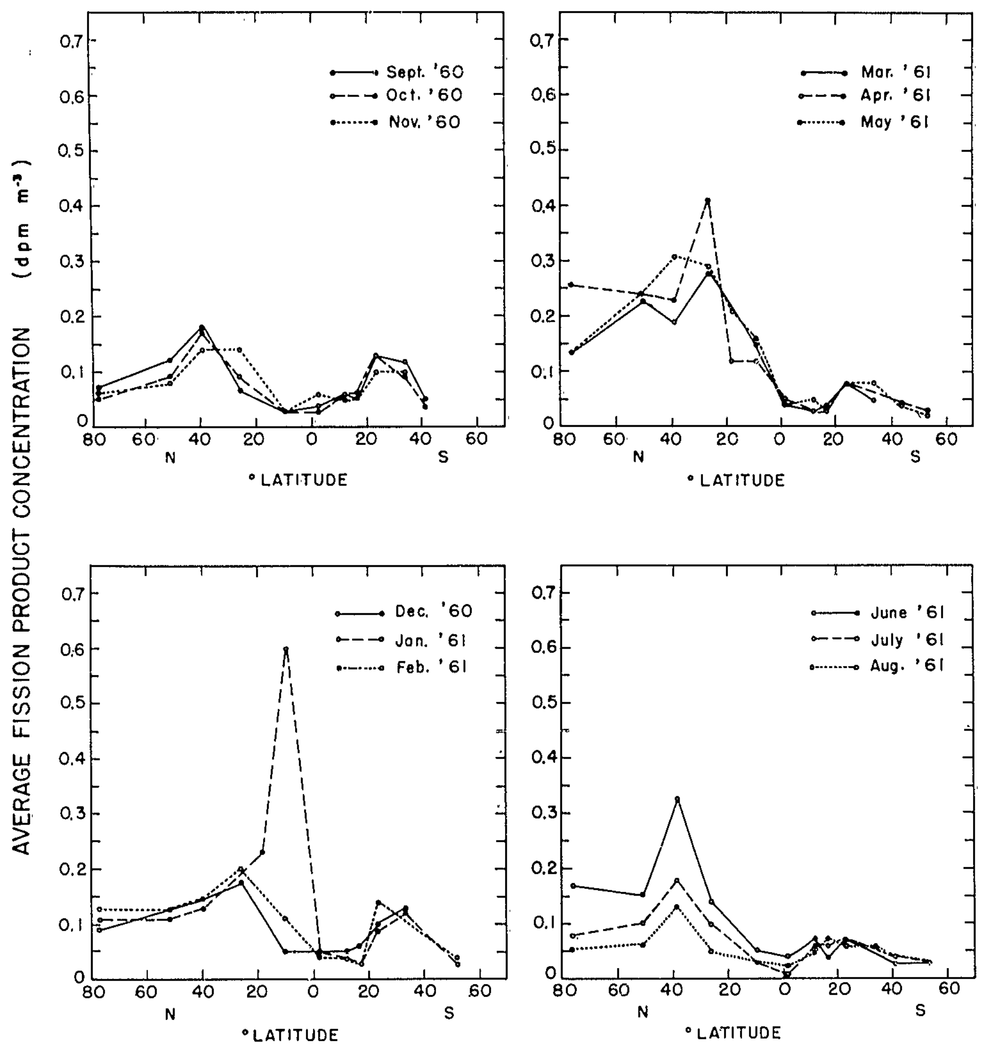
FIGURE 1. (In three sections) Average fission product concentration in surface air along 80° West.

Units are in disintegrations per minute per cubic metre.


AVERAGE FISSION PRODUCT CONCENTRATION (dpm m^{-3})
Sept.'57
Oct.'57
Nov.'57
Dec.'57
Jan.'58
Feb.'58
Mar.'58
Apr.'58
May'58
June'58
July'58
Aug.'58
Sept.'58
Oct.'58
Nov.'58
Dec.'58
Jan.'59
Feb.'59
N
S
° LATITUDE

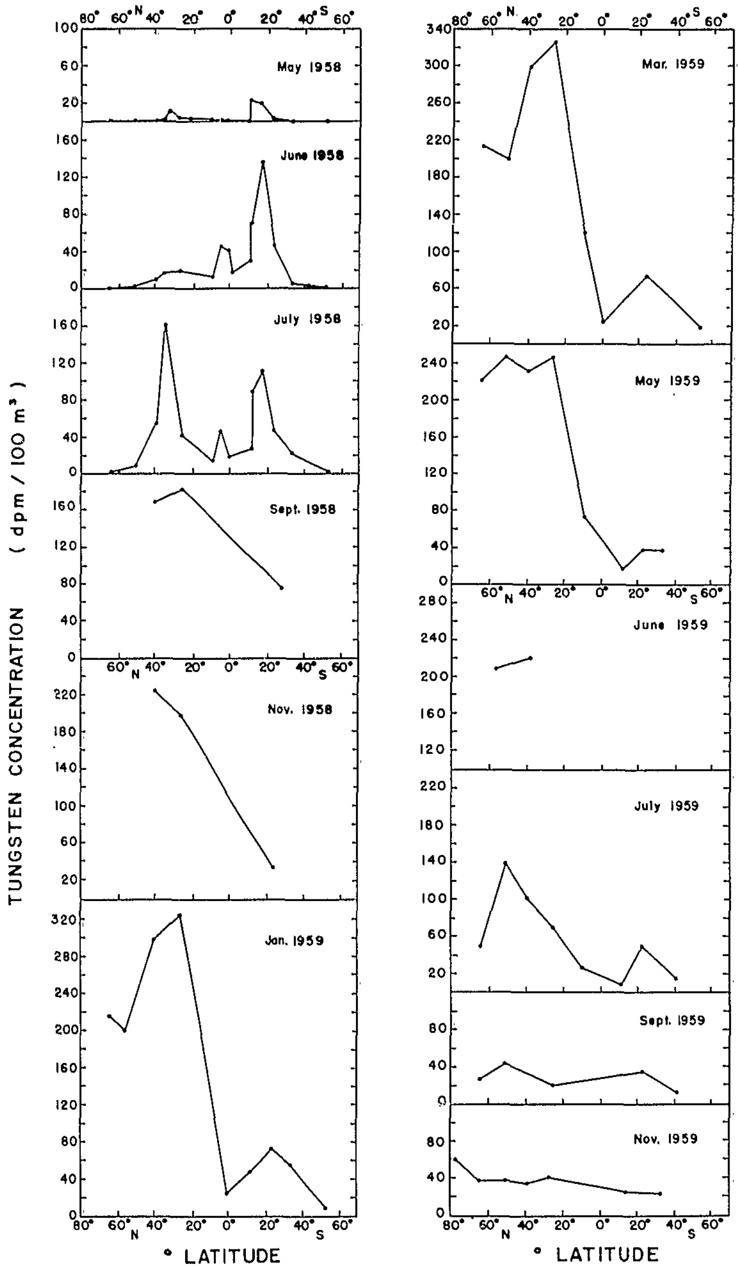

AVERAGE FISSION PRODUCT CONCENTRATION (dpm m^{-3})
Mar.'59
April'59
May'59
Dec.'59
Jan.'60
Feb.'60
3.65
June'59
July'59
Aug.'59
Mar.'60
Apr.'60
May'60
3.23
5.55
Sept.'59
Oct.'59
Nov.'59
June'60
July'60
Aug.'60
N
S
° LATITUDE



AVERAGE FISSION PRODUCT CONCENTRATION (dpm m^{-3})
Sept. '60
Oct. '60
Nov. '60
Mar. '61
Apr. '61
May '61
Dec. '60
Jan. '61
Feb. '61
June '61
July '61
Aug. '61
0.7
0.6
0.5
0.4
0.3
0.2
0.1
0
80 60 40 20 0 20 40 60
N
S
° LATITUDE

FIGURE 2. Average concentration of tungsten 185 in surface air along 80^{o} West.

Units are in disintegrations per minute per 100 cubic metres.


TUNGSTEN CONCENTRATION (dpm / 100 m³)
May 1958
June 1958
July 1958
Sept. 1958
Nov. 1958
Jan. 1959
Mar. 1959
May 1959
June 1959
July 1959
Sept. 1959
Nov. 1959
° LATITUDE
° LATITUDE

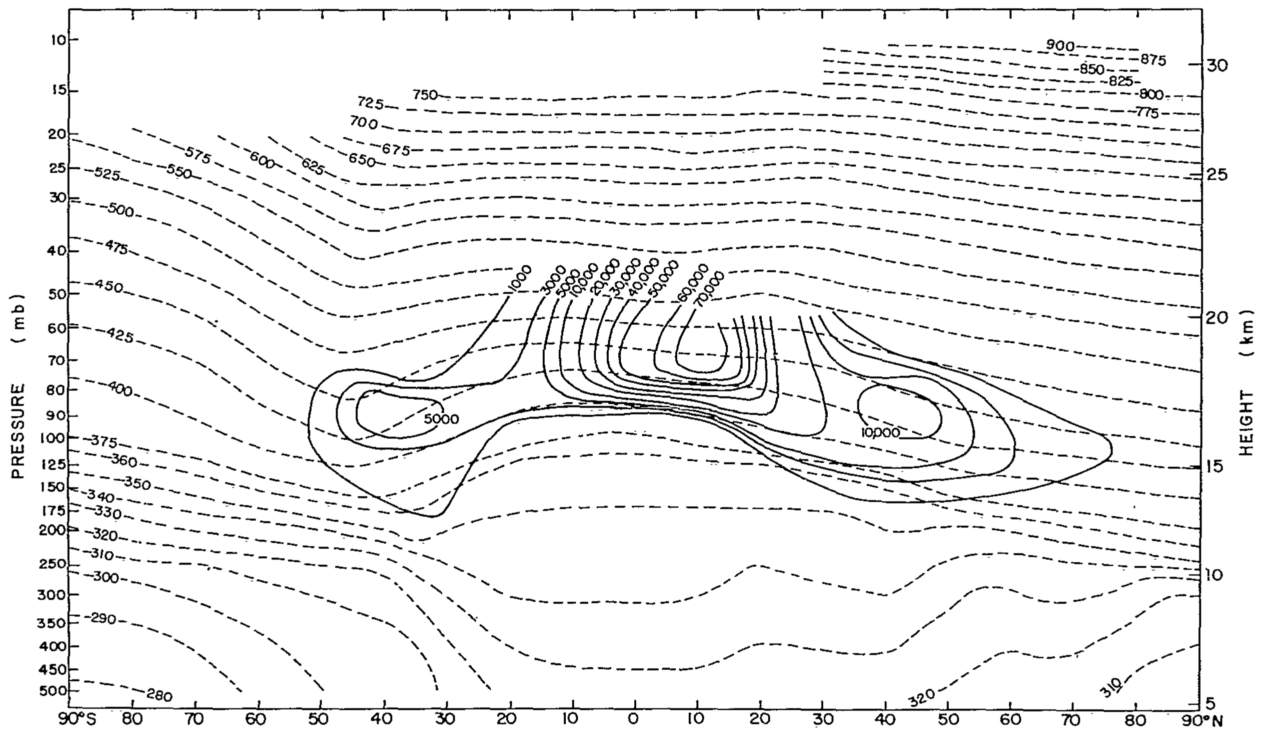


FIGURE 3. Distribution of tungsten 185 (solid lines) and potential temperature (dotted lines) in the stratosphere. Tungsten values for September - October 1958. Units are disintegrations per minute per 1000 standard cubic feet of air. Potential temperatures for July 1957 (degrees Kelvin).

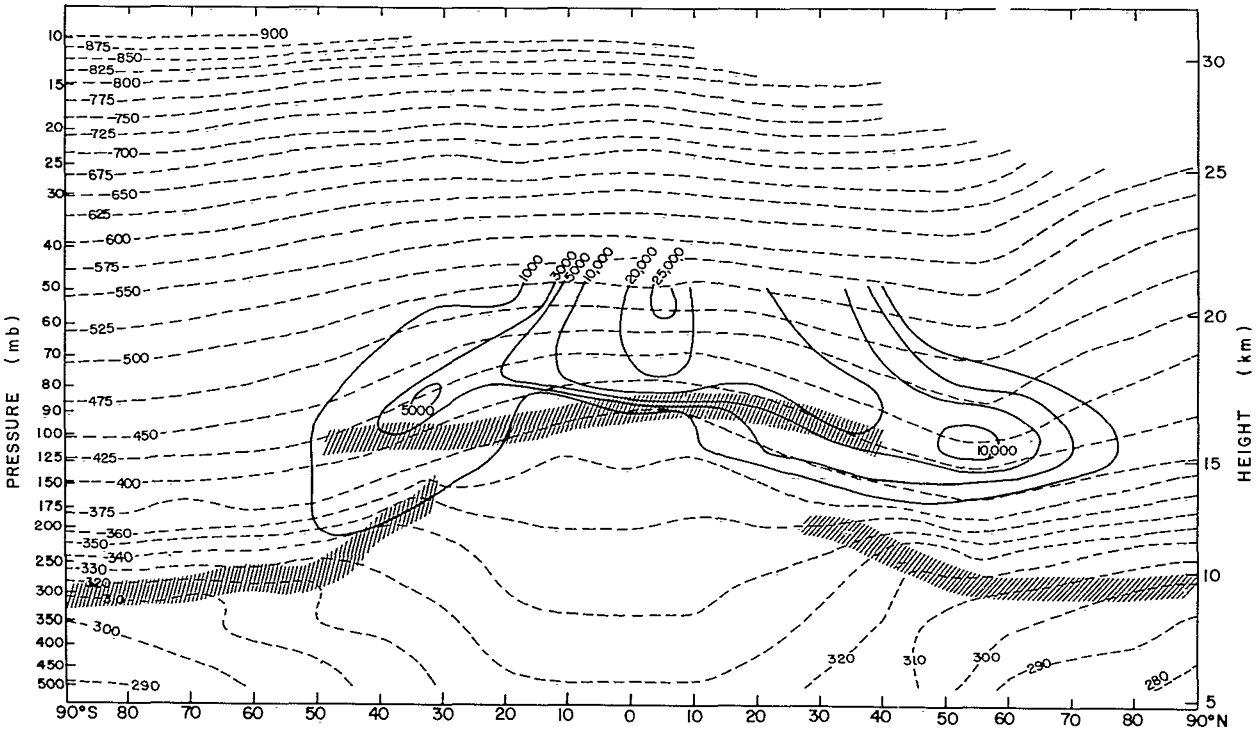


FIGURE 4. Distribution of tungsten 185 (solid lines) and potential temperature (dotted lines) in the stratosphere. Tungsten values for November - December 1958. Potential temperature for December 1957. Shading represents region of tropopause.

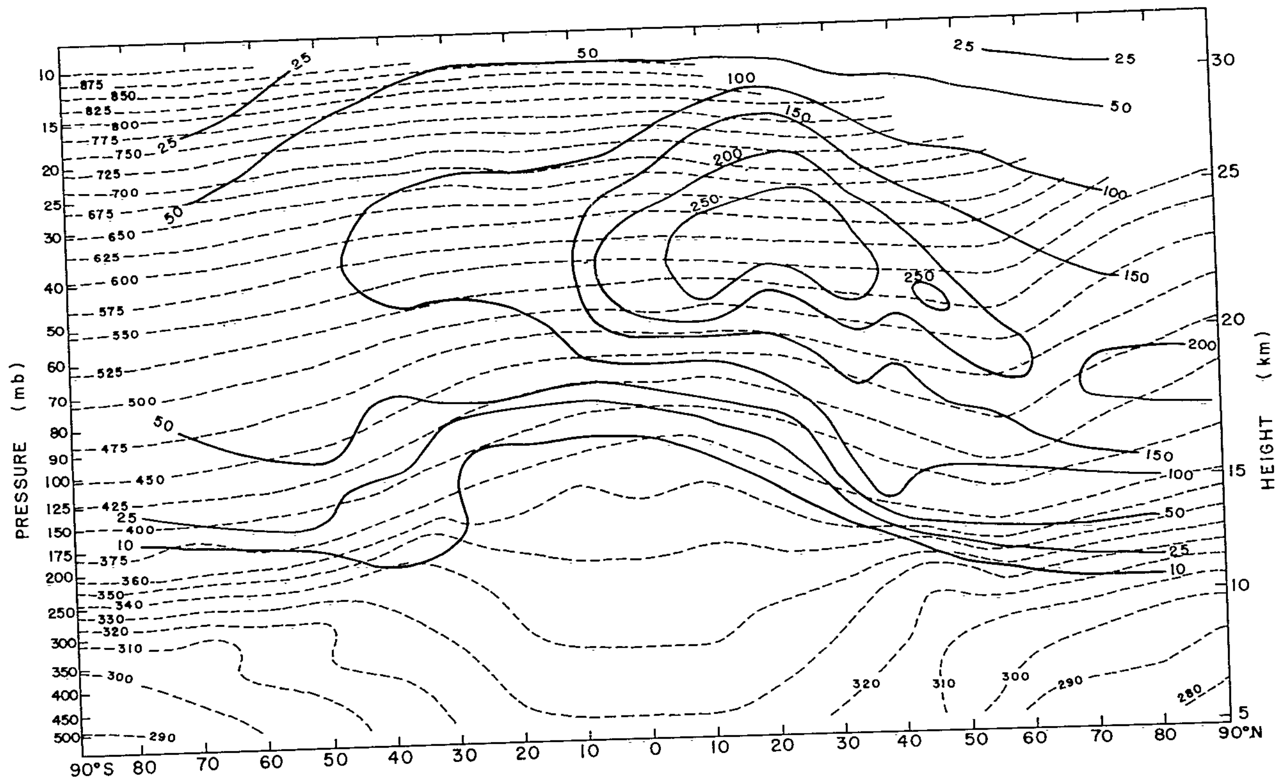


FIGURE 5. Distribution of strontium 90 in stratosphere January - June 1959. Units are disintegrations per minute per 1000 standard cubic feet of air. Potential temperatures for December 1957.

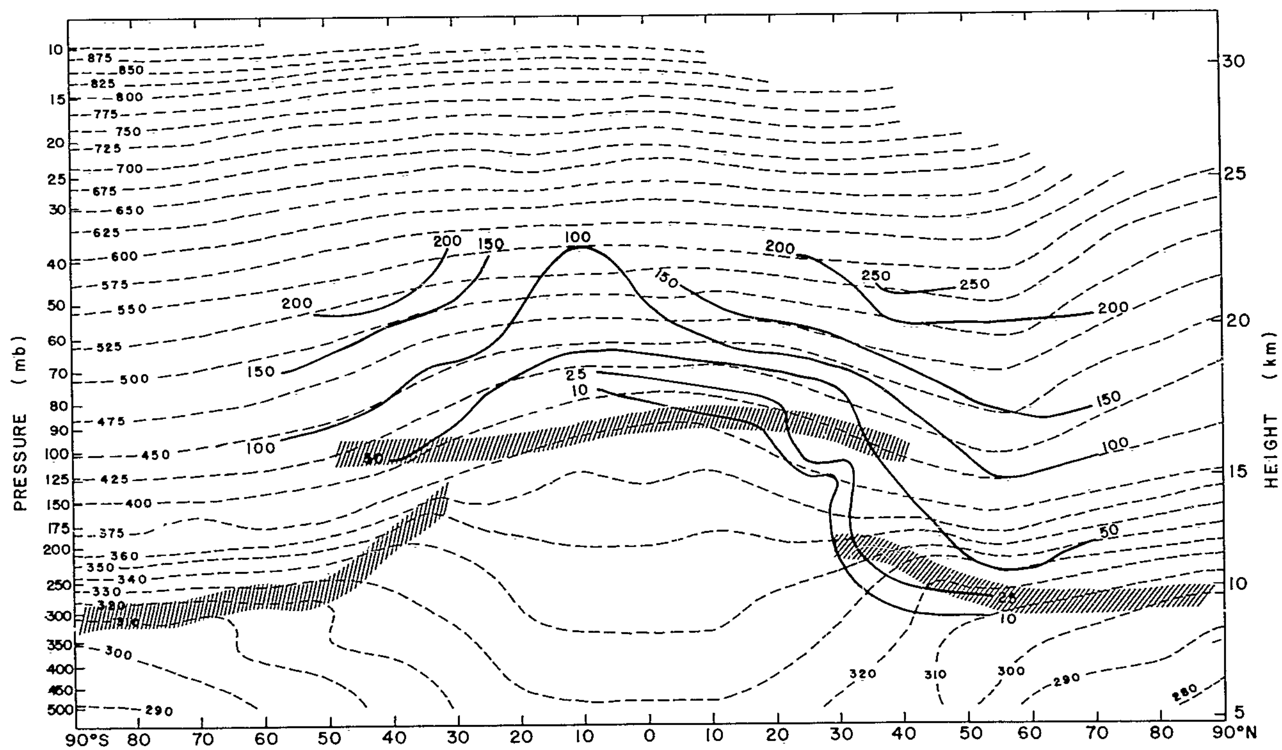


FIGURE 6. Distribution of strontium 90 in the stratosphere January - June 1960. Units are disintegrations per minute per 1000 standard cubic feet of air. Potential temperatures for December 1957.

TABLE 1. Ozone Stations

Station	International Index No.	Location	
Marcus Island, Pacific	91-131	24°17'N	153°58'E
Torishima, Japan	47-963	30°29'N	140°18'E
Tateno, Japan	47-646	36°03'N	140°08'E
Sapporo, Japan	47-412	43°03'N	141°20'E
Washington, D.C., U.S.A	72-405	38°51'N	77°02'W
Abastumani, U.S.S.R.	37-506	41°43'N	42°50'E
(Tbilisi, U.S.S.R.)	37-549	41°41'N	44°57'E
Rome, Italy	16-239	41°48'N	12°36'E
Vladivostok, U.S.S.R.	31-960	43°07'N	131°54'E
Alma Ata, U.S.S.R.	36-870	43°15'N	76°56'E
Green Bay, Wisconsin, U.S.A.	72-645	44°29'N	88°08'W
Bismarck, N. Dakota, U.S.A.	72-764	46°46'N	100°45'W
Arosa, Switzerland	---	46°47'N	09°41'E
(Milano, Italy)	16-080	45°28'N	09°17'E
Caribou, Maine, U.S.A.	72-712	46°50'N	68°00'W
Cambourne, England	03-808	50°13'N	05°19'W
Moosonee, Canada	72-836	51°16'N	80°39'W
Oxford, England	---	51°46'N	01°16'W
(Crawley, England)	03-774	51°05'N	00°13'W
Edmonton, Canada	72-879	53°34'N	113°31'W
Eskdalemuir, Scotland	03-162	55°19'N	03°12'W
(Leuchars, Scotland)	03-171	56°23'N	02°53'W
Aarhus, Denmark	06-070	56°18'N	10°37'E
(Copenhagen, Denmark)	06-180	55°38'N	12°40'E
Uppsala, Sweden	02-076	59°52'N	17°37'E
(Stockholm, Sweden)	02-077	59°21'N	17°57'E
Leningrad, U.S.S.R.	26-063	59°57'N	30°42'E
Lerwick, Scotland	03-005	60°08'N	01°11'W
Reykjavik, Iceland	04-030	64°08'N	21°54'W
(Keflavik, Iceland)	04-018	63°57'N	23°37'W
Resolute, Canada	72-924	74°43'N	94°59'W
Alert, Canada	74-082	82°30'N	62°20'W

TABLE 2. Average ozone amounts
Units: cm of ozone at STP
Number of observations in parentheses

Station	July-September'57		October-December'57		January-March'58	
	$\overline{O}$	$\sigma(O)$	$\overline{O}$	$\sigma(O)$	$\overline{O}$	$\sigma(O)$
Marcus Is.	-	-	-	-	0.241(86)	0.012
Torishima	-	-	0.262(25)	0.012	0.310(74)	0.030
Tateno	0.301(82)	0.022	0.290(85)	0.025	0.356(77)	0.044
Sapporo	-	-	-	-	0.445(29)	0.039
Washington D.C.	0.319(32)	0.020	0.296(31)	0.022	0.358(40)	0.036
Abastumani	0.257(46)	0.018	0.280(37)	0.021	0.347(29)	0.035
Rome	0.315(81)	0.018	0.306(81)	0.028	0.378(90)	0.048
Vladivostok	0.268(22)	0.015	0.259(14)	0.026	-	-
Alma Ata	-	-	0.226(11)	0.021	0.292(13)	0.019
Green Bay	-	-	-	-	0.386(12)	0.028
Bismarck	-	-	-	-	0.375(52)	0.049
Milano	0.303(74)	0.018	0.284(70)	0.029	0.360(64)	0.052
Caribou	-	-	-	-	-	-
Cambourne	0.301(84)	0.029	0.277(77)	0.030	0.352(57)	0.035
Moosonee	0.325(75)	0.024	0.319(72)	0.031	0.397(83)	0.046
Crawley	0.331(91)	0.030	0.288(86)	0.036	0.368(78)	0.054
Edmonton	0.304(76)	0.027	0.291(74)	0.044	0.379(35)	0.062
Leuchars	0.325(5)	0.024	0.294(64)	0.029	0.380(62)	0.073
Copenhagen	0.352(91)	0.026	0.306(85)	0.033	0.428(78)	0.073
Stockholm	0.318(73)	0.017	0.275(71)	0.030	0.404(68)	0.074
Leningrad	0.347(57)	0.025	0.303(10)	0.019	0.460(24)	0.023
Lerwick	0.330(83)	0.032	0.280(46)	0.026	0.405(56)	0.049
Keflavik	0.307(73)	0.022	0.291(71)	0.032	0.401(72)	0.075
Tromso	0.302(80)	0.018	0.277(86)	0.040	0.420(86)	0.093
Resolute	0.310(57)	0.040	0.281(20)	0.033	0.442(8)	0.049
Alert	0.301(70)	0.020	-	-	0.433(8)	0.116

April-June'58		July-September'58		October-December'58	
$\overline{O}$	$\sigma(O)$	$\overline{O}$	$\sigma(O)$	$\overline{O}$	$\sigma(O)$
0.287(80)	0.019	0.282(91)	0.012	0.257(90)	0.013
0.341(76)	0.026	0.300(84)	0.024	0.256(83)	0.016
0.355(81)	0.026	0.302(84)	0.021	0.290(70)	0.023
0.422(69)	0.044	0.326(82)	0.028	0.352(70)	0.049
0.352(77)	0.052	0.312(75)	0.037	0.301(86)	0.025
0.324(24)	0.035	0.276(40)	0.017	0.275(44)	0.028
0.374(90)	0.040	0.311(91)	0.017	0.311(79)	0.032
0.322(22)	0.038	0.273(24)	0.014	0.272(37)	0.045
0.283(22)	0.022	0.254(36)	0.024	0.252(15)	0.032
0.356(66)	0.028	0.295(64)	0.023	0.313(37)	0.048
0.369(81)	0.046	0.305(87)	0.045	0.300(71)	0.033
0.366(66)	0.034	0.312(83)	0.026	0.294(66)	0.028
0.376(57)	0.031	0.321(82)	0.035	0.318(78)	0.046
0.368(89)	0.032	0.320(68)	0.031	0.296(63)	0.050
0.413(91)	0.050	0.339(92)	0.028	0.359(89)	0.050
0.394(87)	0.028	0.331(90)	0.033	0.299(71)	0.039
0.380(90)	0.041	0.317(91)	0.027	0.334(91)	0.042
0.381(68)	0.032	0.328(85)	0.031	0.308(78)	0.040
0.414(88)	0.038	0.337(90)	0.033	0.308(58)	0.032
0.374(82)	0.050	0.309(81)	0.033	0.283(56)	0.042
0.397(54)	0.039	0.337(54)	0.031	0.272(15)	0.018
0.389(91)	0.037	0.319(92)	0.032	0.277(50)	0.034
0.385(87)	0.036	0.316(92)	0.038	0.319(56)	0.047
0.396(85)	0.039	0.299(80)	0.027	0.290(76)	0.046
0.406(68)	0.047	0.333(92)	0.024	0.383(26)	0.079
0.416(77)	0.065	0.310(79)	0.033	0.313(2)	0.013

TABLE 3. Meridional flux of ozone by transient eddies

Units are cm of ozone at STP m sec^{-1}.
(Number of cases in parentheses)

Latitude Belts	July-Sep 1957	Oct-Dec 1957	Jan-Mar 1958	Apr-June 1958	July-Sep 1958	Oct-Dec 1958
100 mb						
Marcus Island 24°N	-	-	-0.0111 (23)	-0.0316 (68)	+0.0022 (81)	-0.0442 (72)
Japan (3 stations)	-0.0664 (172)	-0.0050 (228)	-0.0558 (377)	+0.0183 (404)	-0.0621 (511)	-0.0572 (459)
38° -45°N (6 stations)	+0.0126 (161)	-0.0011 (184)	+0.0813 (232)	+0.0133 (419)	+0.0064 (398)	-0.0035 (332)
45° -55°N (7 stations)	+0.0519 (568)	+0.0510 (529)	+0.0725 (499)	+0.0043 (738)	+0.0340 (749)	+0.0461 (606)
55° -60°N (5 stations)	+0.0346 (310)	+0.0819 (261)	+0.0777 (380)	-0.0103 (548)	+0.0540 (546)	+0.0446 (345)
Keflavik 64°N	+0.0381 (140)	+0.0240 (125)	-0.1779 (136)	+0.0736 (167)	-0.0081 (166)	-0.1968 (105)
Resolute 75°N	+0.0647 (29)	+0.1022 (18)	-0.0421 (6)	+0.0383 (99)	+0.0442 (121)	+0.1111 (18)
Alert 82.5°N	+0.0064 (120)	+0.0049 (13)	-0.9798 (6)	+0.0183 (124)	+0.0220 (128)	+0.0752 (10)
50 mb						
Marcus Island 24°N	-	-	+0.0313 (12)	+0.0021 (55)	+0.0006 (75)	+0.0140 (60)
Japan (3 stations)	-0.0075 (148)	-0.0031 (211)	-0.0416 (312)	+0.0183 (319)	-0.0148 (461)	+0.0126 (393)
38° -45°N (6 stations)	+0.0038 (79)	+0.0228 (94)	+0.1181 (166)	+0.0054 (321)	+0.0050 (302)	-0.0261 (220)
45° -55°N (7 stations)	+0.0319 (409)	+0.0106 (393)	+0.1151 (372)	+0.0099 (568)	+0.0222 (592)	+0.0162 (638)
55° -60°N (5 stations)	+0.0172 (180)	+0.0683 (151)	+0.0767 (192)	+0.0101 (359)	+0.0356 (335)	+0.0952 (207)
Keflavik 64°N	+0.0169 (141)	+0.0572 (117)	-0.2680 (125)	+0.0390 (166)	+0.0030 (161)	-0.2420 (96)
Resolute 75°N	+0.0406 (25)	-0.1011 (13)	-0.0578 (4)	+0.0077 (87)	+0.0341 (110)	+0.2015 (14)
Alert 82.5°N	+0.0022 (113)	+0.0675 (5)	-0.4754 (7)	-0.0186 (126)	+0.0087 (121)	+0.0833 (9)

TABLE 4. Meridional flux of ozone by standing eddies
Units are cm of ozone at STP m sec^{-1}.

	July-Sept. 1957	Oct.-Dec. 1957	Jan-March 1958	April-June 1958	July-Sept. 1958	Oct-Dec. 1958
100 mb	-0.0070	+0.0162	+0.0143	+0.0048	-0.0087	+0.0300
50 mb	+0.0038	+0.0123	+0.0439	+0.0078	-0.0110	+0.0256

TABLE 5. Ozone budget

(Units are in cm^3 of ozone at STP sec^{-1} when multiplied by 10^9)

	July-Sept. 1957	Jan.-March 1958	July-Sept. 1958
Transport of ozone across 50°N			
a) by transient eddies	+2.9	+10.4	+2.0
b) by standing eddies	+0.3	+ 4.0	-1.0
c) by mean meridional motions	-0.9	- 1.9	?
Transport from content change north of 50°N	-5.2	+ 9.0	-5.2
Transport of ozone across 40°N			
a) by transient eddies	+0.4	+12.6	+0.5
b) by standing eddies	+0.4	+ 5.0	-1.2
c) by mean meridional motions	-2.3	- 3.8	?
Transport from content change north of 40°N	-7.4	+11.6	-7.4
Tropospheric downward flux	$\sim$ 6		

TABLE 6. Terminal velocity of spherical particles
Units: cm sec^{-1}

Altitude (Km)	radius 0.01 μ	0.03 μ	0.1 μ	0.2 μ	1.0 μ
30	0.0022	0.0067	0.021	0.067	0.24
25	0.0010	0.0030	0.010	0.032	0.13
20	0.00048	0.0015	0.005	0.017	0.06
15	0.00022	0.00065	0.0023	0.008	0.048
10	0.00010	0.00031	0.0013	0.0048	0.038

TABLE 7. Distance fallen in two months at terminal velocity
Units: meters

Altitude (Km)	radius 0.01 μ	0.03 μ	0.1 μ	0.3 μ	1.0 μ
30	114	347	1089	3473	12442
25	52	156	518	1659	6739
20	25	78	259	881	3110
15	11	34	119	415	2488
10	5	16	67	249	1970

TABLE 8. Latitude of maximum concentration of tungsten.

	Latitude
September - October 1958	10° N
November - December 1958	4° N
January - February 1959	5° S
March - April 1959	7° N
May - June 1959	0° N
July - August 1959	0° N
September - October 1959	5° N
November - December 1959	5° N
January - February 1960	5° N
March - April 1960	?
May - June 1960	6° N

TABLE 9. Covariance of meridional velocity and temperature at ozone stations.

Units are ^{o}C m sec^{-1} (No. of cases in parentheses)

Latitude Belts	July-Sep 1957	Oct-Dec 1957	Jan-Mar 1958	Apr-June 1958	July-Sep 1958	Oct-Dec 1958
50 mb						
Marcus Island	--	--	--	--	--	--
Japan 30-43°N	+0.39	+2.84	+2.94	+2.98	+0.18	+2.18
(3 stations)	(404)	(476)	(427)	(351)	(480)	(442)
38°-45°N	+0.62	+3.88	+5.54	+2.23	+3.25	+3.35
(6 stations)	(405)	(320)	(388)	(459)	(416)	(279)
45°-55°N	+3.19	+1.88	+5.75	+2.64	+2.56	+5.88
(7 stations)	(583)	(583)	(640)	(651)	(639)	(656)
55°-60°N	+4.37	+12.47	+11.25	+8.37	+5.58	+7.12
(5 stations)	(289)	(209)	(223)	(384)	(355)	(331)
Keflavik 64°N	+0.70	+6.24	-40.85	+0.62	-1.23	-20.82
	(177)	(149)	(149)	(174)	(160)	(151)
Resolute 75°N	+3.42	-11.69	+0.78	+6.87	-1.23	+17.82
	(38)	(42)	(42)	(116)	(109)	(68)
Alert 82.5°N	-1.42	-10.34	-31.85	+0.11	-2.46	-18.89
	(142)	(106)	(124)	(156)	(137)	(127)

TABLE 10. Covariance of meridional and vertical velocities.

Northern hemisphere, January 1958. Transient eddy effects.

Units: $cm^2\ sec^{-2}$.

Pressure layer (mb)	Latitude						
	20°	30°	40°	50°	60°	70°	80°
1000 - 850	+109	+366	+350	+197	+ 92	+150	- 83
850 - 700	+ 86	+324	+365	+247	+ 96	+196	+ 4
700 - 500	+ 74	+354	+463	+365	+154	+233	+133
500 - 300	+261	+1192	+906	+343	+317	+725	+183
300 - 200	+246	+697	+844	+494	+475	+350	+ 58
200 - 100	+240	+232	+167	+258	+367	+167	+100
100 - 50	- 35	- 97	-103	- 11	+142	+108	+100

TABLE 11. Covariance between ozone amount and temperature

Units are cm of ozone at STP °C
(Number of cases in parentheses)

Latitude Belts	July-Sep 1957	Oct-Dec 1957	Jan-March 1958	Apr-June 1958	July-Sep 1958	Oct-Dec 1958
50 mb						
Japan 30-43°N (3 stations)	+0.0033 (148)	+0.0433 (218)	-0.0091 (324)	+0.0358 (373)	+0.0172 (462)	+0.0296 (408)
38° - 45°N (6 stations)	+0.0044 (299)	+0.0128 (192)	+0.0858 (243)	+0.0344 (472)	+0.0273 (479)	+0.0081 (331)
45° - 55°N (7 stations)	+0.0408 (596)	+0.0286 (577)	+0.1182 (542)	+0.0043 (843)	+0.0533 (913)	+0.0605 (851)
55° - 60°N (5 stations)	+0.0320 (356)	+0.0382 (281)	+0.1535 (290)	-0.0292 (454)	+0.0896 (499)	+0.0652 (348)
Keflavik 64°N	+0.0663 (144)	+0.0764 (119)	+0.4840 (129)	+0.1034 (168)	+0.0573 (174)	+0.0703 (100)
Resolute 75°N	+0.0852 (93)	-0.0063 (21)	+0.2295 (10)	-0.1036 (115)	+0.0215 (171)	+0.5065 (42)
Alert 82.5°N	+0.0465 (126)	-0.1245 (5)	+1.3418 (8)	-0.2246 (141)	+0.0407 (145)	+0.1250 (10)

TABLE 12. Zonal flux of ozone by transient eddies
Units are cm of ozone at STP m sec^{-1}
(Number of cases in parentheses)

Latitude Belts	July-Sep 1957	Oct-Dec 1957	Jan-Mar 1958	Apr-June 1958	July-Sep 1958	Oct-Dec 1958
50 mb						
Marcus Island	-	-	+0.0485 (12)	-0.1135 (55)	-0.0054 (75)	-0.0357 (60)
Japan 30-43°N (3 stations)	-0.0095 (148)	+0.0501 (211)	+0.0066 (312)	+0.0898 (319)	-0.0472 (461)	+0.0544 (393)
38° -45°N (6 stations)	-0.0412 (79)	+0.0733 (94)	-0.0452 (166)	+0.0976 (321)	-0.0241 (300)	+0.0249 (220)
45° -55°N (7 stations)	-0.0317 (409)	+0.0827 (388)	-0.1429 (372)	+0.0825 (531)	-0.0378 (592)	+0.0550 (518)
55° -60°N (5 stations)	-0.0292 (180)	+0.0506 (151)	-0.1727 (192)	+0.0924 (359)	-0.0173 (335)	+0.0201 (207)
Keflavik 64°N	-0.0328 (141)	-0.0055 (117)	-0.5801 (125)	+0.1550 (166)	-0.0351 (161)	-0.1132 (96)
Resolute 75°N	-0.0582 (25)	-0.0670 (13)	+0.0819 (4)	+0.1223 (87)	+0.0050 (110)	-0.5867 (14)
Alert 82.5°N	-0.0248 (113)	+0.1663 (5)	+0.1221 (7)	-0.0270 (126)	-0.0270 (121)	+0.1174 (9)

TABLE 13. Mean zonal and meridional components of Rocket Network winds.

Units: knots

Height		Ft. Churchill 58.8°N, 94.3°W				Tonopah Range Nev. 38.0°N, 116.5°W				Wallops Is, Va. 37.8°N, 75.5°W				Pt. Mugu. Cal. 34.1°N, 119.1°W			
Km	Ft	$\bar{u}$	$\bar{v}$	(N)	$\sigma(v)$	$\bar{u}$	$\bar{v}$	(N)	$\sigma(v)$	$\bar{u}$	$\bar{v}$	(N)	$\sigma(v)$	$\bar{u}$	$\bar{v}$	(N)	$\sigma(v)$
61.0	200,000	+62.0	+12.0	(1)		+53.5	-5.00	(2)		+100.2	+7.1	(14)	19.2	+105.9	+15.0	(9)	
54.9	180,000	+110.5	-0.75	(4)		+48.5	+19.0	(2)		+82.2	+14.8	(29)	21.6	+90.3	+18.9	(26)	26.2
48.8	160,000	+94.6	-1.7	(7)		+25.0	+5.0	(1)		+77.0	+15.5	(38)	22.0	+78.8	+10.6	(39)	17.1
42.7	140,000	+46.5	-13.6	(15)	18.7	+68.5	+11.5	(15)	17.0	+61.4	+14.0	(47)	21.9	+65.7	+2.1	(44)	17.0
36.6	120,000	+34.8	-13.5	(24)	28.1	+26.7	-2.4	(36)	12.6	+43.8	+1.3	(51)	17.5	+44.0	+5.7	(55)	12.0
30.5	100,000	+17.2	-9.3	(30)	23.9	+9.6	-0.47	(36)	10.9	+23.6	+5.4	(57)	10.7	+22.2	+3.4	(63)	10.7
24.4	80,000	+16.9	-10.1	(33)	17.0	-0.23	-2.3	(30)	6.8	+7.3	+2.5	(42)	7.6	+5.5	+1.2	(60)	7.9
18.3	60,000	+14.3	-5.3	(24)	10.2	+10.7	-2.4	(16)	9.3	+23.4	-2.2	(31)	13.7	+19.4	-0.39	(49)	9.0

TABLE 13 (cont.). Mean zonal and meridional components of Rocket Network winds

Units: knots.

Height		33.3°N, 106.5°W				32.9°N, 106.1°W				30.5°N, 86.5°W				28.2°N, 80.6°W			
		White Sands, N. Mex.				Holloman, N. Mex.				Eglin Field, Fa.				Cape Canaveral, Fla.			
Km	Ft	$\bar{u}$	$\bar{v}$	(N)	$\sigma(v)$	$\bar{u}$	$\bar{v}$	(N)	$\sigma(v)$	$\bar{u}$	$\bar{v}$	(N)	$\sigma(v)$	$\bar{u}$	$\bar{v}$	(N)	$\sigma(v)$
61.0	200,000	+107.2	+23.8	(32)	24.5	+113.3	+28.3	(7)		+84.0	-0.27	(11)	16.0	+112.3	+7.5	(6)	
54.9	180,000	+103.6	+28.7	(40)	24.1	+60.6	+17.0	(8)		+99.8	+9.8	(12)	16.2	+114.7	+32.4	(11)	22.1
48.8	160,000	+96.0	+19.2	(40)	23.9	+62.0	+ 0.17	(12)	25.3	+109.4	+14.9	(14)	19.2	+83.4	+30.3	(17)	28.9
42.7	140,000	+79.6	+2.7	(43)	16.0	+54.0	-8.1	(16)	28.3	+93.7	+8.5	(15)	23.3	+68.5	+1.8	(16)	20.4
36.6	120,000	+47.7	-0.25	(56)	10.5	+33.0	-2.4	(19)	15.3	+71.9	-5.6	(14)	10.5	+52.4	+0.10	(21)	14.7
30.5	100,000	+16.6	+1.9	(59)	9.9	+8.1	-4.6	(19)	8.4	+32.1	+7.0	(11)	6.0	+22.0	+2.3	(21)	8.5
24.4	80,000	+8.6	+0.02	(62)	7.4	+17.5	-6.2	(4)		+5.0	-8.0	(1)		+10.9	+1.6	(12)	5.2
18.3	60,000	+21.6	-1.3	(45)	9.2	+20.2	+4.0	(4)						+37.2	+1.8	(6)	

TABLE 14. Angular momentum transport from Rocket Network winds

Units: $knot^2$

Height Km	Ft		58.8°N, 94.3°W Ft. Churchill		38.0°N, 116.5°W Tonopah Range Nev.		37.8°N, 75.5°W Wallops Is. Va.		34.1°N, 119.1°W Pt. Mugu. Cal.	
61.0	200,000	$\bar{u}\,\bar{v}$					+708.6	(14)		
		$\overline{u'v'}$					+257.4			
54.9	180,000	$\bar{u}\,\bar{v}$					+1212.8	(29)	+1708.9	(26)
		$\overline{u'v'}$					+362.2		+213.5	
48.8	160,000	$\bar{u}\,\bar{v}$					+1193.9	(38)	+835.0	(39)
		$\overline{u'v'}$					+385.6		-95.3	
42.7	140,000	$\bar{u}\,\bar{v}$	-632.9	(15)	+789.6	(15)	+858.9	(47)	+137.3	(44)
		$\overline{u'v'}$	+183.5		+176.4		+134.1		+70.5	
36.6	120,000	$\bar{u}\,\bar{v}$	-468.8	(24)	-65.3	(36)	+55.8	(57)	+253.0	(55)
		$\overline{u'v'}$	+169.8		+264.8		+342.6		+24.0	
30.5	100,000	$\bar{u}\,\bar{v}$	-160.5	(30)	-4.5	(36)	+128.2	(57)	+74.9	(63)
		$\overline{u'v'}$	+187.4		+133.6		+73.7		+4.7	
24.4	80,000	$\bar{u}\,\bar{v}$	-170.9	(33)	+0.54	(30)	+18.3	(42)	+6.3	(60)
		$\overline{u'v'}$	56.3		+42.5		+8.4		+68.3	
18.3	60,000	$\bar{u}\,\bar{v}$	-75.8	(24)	-26.1	(16)	-50.5	(60)	-7.5	(49)
		$\overline{u'v'}$	-41.2		+29.1		-51.2		+7.3	

TABLE 14. (cont.) Angular momentum transport from Rocket Network winds

Units: $knot^2$.

Height Km	Height Ft		33.3°N, 106.5°W White Sands, N. Mex.		32.9°N, 106.1°W Holloman, N. Mex.		30.5°N, 86.5°W Eglin Field, Fla.		28.2°N, 80.6°W Cape Canaveral, Fla.	
61.0	200,000	$\bar{u}\,\bar{v}$	+2555.0	(32)			-22.9	(11)		
		$\overline{u'v}$	+524.8				+48.5			
54.9	180,000	$\bar{u}\,\bar{v}$	+2974.8	(40)			+980.9	(12)		
		$\overline{u'v'}$	+401.0				-179.4			
48.8	160,000	$\bar{u}\,\bar{v}$	+1846.1	(40)	+10.3	(12)	+1632.5	(14)	+2525.1	(17)
		$\overline{u'v'}$	+459.5		-404.8		-161.9		+177.8	
42.7	140,000	$\bar{u}\,\bar{v}$	+218.4	(43)	-435.4	(16)	+800.0	(15)	+124.2	(16)
		$\overline{u'v'}$	+115.2		+299.8		+354.3		+79.3	
36.6	120,000	$\bar{u}\,\bar{v}$	-11.9	(56)	-79.9	(19)	-405.5	(14)	+5.0	(21)
		$\overline{u'v'}$	+37.2		+408.7		+86.8		+16.7	
30.5	100,000	$\bar{u}\,\bar{v}$	+32.4	(59)	-37.1	(19)	+224.6	(11)	+50.2	(21)
		$\overline{u'v'}$	-4.4		+38.9		+132.0		+62.5	
24.4	80,000	$\bar{u}\,\bar{v}$	+0.14	(62)						
		$\overline{u'v'}$	+49.6							
18.3	60,000	$\bar{u}\,\bar{v}$	-28.3	(45)						
		$\overline{u'v'}$	+19.4							

THE GENERAL CIRCULATION OF THE ATMOSPHERE AND ITS EFFECTS ON THE MOVEMENT OF TRACE SUBSTANCES

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This paper was presented at the Symposium on Trace Gases and Natural and Artificial Radioactivity in the Atmosphere held at Utrecht, The Netherlands 8-14 August, 1962. The Symposium was sponsored by the Commission on Atmospheric Chemistry and Radioactivity of the International Association for Meteorology and Atmospheric Physics of the International Union for Geodesy and Geophysics and the World Meteorological Organization.

ABSTRACT

A brief summary is made of present knowledge of horizontal and vertical motions in the atmosphere with particular reference to large scale motions which may be expected to redistribute trace substances over long time periods. Mean motions, standing eddy motions and transient eddy motions are defined and discussed separately, with variance data being used as a measure of the transport properties of the eddies. Hemispheric data concerning both horizontal and vertical components are available up to about 25 km, with limited samples of horizontal component data to 60 km. Indications are that eddy motions, rather than mean motions, are the main transport process up to 25 km. Variance data show maxima in both horizontal and vertical components in the vicinity of the tropospheric jet stream and in the vicinity of the polar night stratospheric jet. In the troposphere the smallest values of both horizontal and vertical variances occur in the low latitudes of the tropics. The data are applied to a discussion of the distribution of long lived fission products in the atmosphere. A possible explanation for the inclination of the concentration isolines of the trace substances to the isentropic surfaces is presented. Finally, the general approach is applied to the particular case of the budget of ozone in the stratosphere.

1. Introduction

The ultimate aim of much of the work presented at the Symposium is to achieve a complete understanding of the distribution of trace substances in the atmosphere. To do this it is necessary to consider the information about the sources and sinks in conjunction with that about the meteorological structure. It is also possible to approach the problem from another point of view and try to deduce information about the meteorological motions from the observed distributions of trace substances but here again the deductions must be compatible with the general body of knowledge about the meteorological structure. The two points of view are actually complementary; for example, while it has been possible to construct from meteorological observations budgets representing the flow of angular momentum and energy within the atmosphere the same observations do not readily lend themselves to the construction of a mass budget, whereas observations of trace substances may allow this to be done. Furthermore, while the network of wind reporting stations and the frequency of observation may limit the time and space scales of the atmospheric motions observed, nevertheless the distribution of the trace substances represents the resultant of all motions.

In the present contribution it is intended to review the current knowledge concerning atmospheric motions that appears pertinent to

the problems perculiar to the trace substances. The general approach to the construction of a global budget for a particular trace substance is first formulated and later illustrated by ref erence to atmospheric ozone. Throughout the emphasis will be upon the global space scale and upon the time scale of months or years. The information discussed is the work of a large number of individuals many of whom have worked under the guidance of Professor V.P. Starr during the past 14 years. Much of the data concerning the troposphere has been published elsewhere and although discussed at the meeting will not be reiterated in the conference proceedings. The climatology of the stratosphere is presently under investigation at the Massachusetts Institute of Technology and the figures to be quoted for this region represent preliminary values for the first 12 months of the IGY period. The final values will be published later by the individuals concerned. I am extremely grateful to my colleagues Dr. A. A. Barnes Jr., Mr. A.H. Oort and Mr. R. E. Dickinson for access to their unpublished work.

2. The General Problem of Transport

Suppose that Q represents a certain property of the air (for example, the concentration of a trace substance) at a particular instant of time and at a point in the atmosphere and V represents the northward component of the wind at the same point and time. Following the

practice common in general circulation studies Q and V may be resolved as follows:

$$Q = \overline{Q} + Q'$$
$$V = \overline{V} + V'$$

where the bars denote a time average (for example over a season) and the primes denote instantaneous departures from the averages.

The instantaneous northward transport of Q is

$$QV = \overline{Q}\,\overline{V} + Q'V' + \overline{Q}\,V' + Q'\,\overline{V}$$

and the time averaged northward transport at the particular point is

$$\overline{QV} = \overline{Q}\,\overline{V} + \overline{Q'V'} \qquad (1)$$

Likewise the time averaged eastward transport is given by

$$\overline{Qu} = \overline{Q}\,\overline{u} + \overline{Q'u'} \qquad (2)$$

The average northward transport of Q past a fixed point, as given by equation (1), consists of two parts; there is transport due to the mean meridional motion at the point and transport by transient eddy processes. The second type of flux arises if there is a correlation between the meridional component of the wind and Q (such as might arise, for example, if northward moving parcels of air contained a higher concentration of a trace substance than southward moving parcels).

When the analysis is extended to include the northward flux across an entire latitude circle instead of at a point then another term may arise due to the presence of a systematic relationship

between Q and V at a series of points around the latitude circle. If we take the latitudinal average (denoted by square brackets) of Equation (1) then

$$\left[\overline{QV}\right] = \left[\bar{Q}\,\bar{V}\right] + \left[\overline{Q'V'}\right]$$

Further

$$\bar{Q} = \left[\bar{Q}\right] + \bar{Q}^{*}$$

$$\bar{V} = \left[\bar{V}\right] + \bar{V}^{*}$$

where $\bar{Q}^{*}$ and $\bar{V}^{*}$ represents deviations from the average value for a particular latitude circle.

Hence

$$\left[\bar{Q}\,\bar{V}\right] = \left[\bar{Q}\right]\left[\bar{V}\right] + \left[\bar{Q}^{*}\,\bar{V}^{*}\right]$$

and

$$\left[\overline{QV}\right] = \left[\bar{Q}\right]\left[\bar{V}\right] + \left[\bar{Q}^{*}\,\bar{V}^{*}\right] + \left[\overline{Q'V'}\right] \tag{3}$$

The time average transport over a complete latitude circle at a particular height thus consists of three terms, transport by a mean meridional motion $\left[\bar{V}\right]$, transport by so called standing eddies $\left[\bar{Q}^{*}\,\bar{V}^{*}\right]$ and transport by transient eddies. A similar approach may be taken to the vertical flux.

This type of resolution has been used in turbulence studies since the time of Osborne Reynolds. It has been applied extensively to

the general circulation of the atmosphere since 1949 (for example by Priestley, (1949), Widger (1949) and Starr (1954)). The early studies were in connection with the sensible heat and momentum budgets of the atmosphere. They were later extended to include fluxes of latent heat, water vapour, and ozone.

Thus in order to obtain a true measure of the flux of a trace substance it is necessary to know its concentration as a function of height at a number of stations round the globe together with concomitant wind information. Apart from water vapour, which has been discussed by Peixoto (1958), the only trace substance which may be treated by these techniques is ozone and the ozone results will be examined later in the paper.

It will be recalled that equations like (3) have been investigated for the cases where Q represents zonal momentum and energy (Starr, 1954). The results for momentum showed that transient eddies played a major role in the budget with standing eddies providing a northward transport in middle latitudes of the northern hemisphere equal to about one-third of that provided by the transient eddies. The northward transport of heat was also accomplished principally by the transient eddies. It has always proved extremely difficult to measure the mean motion transport as $\overline{V}$ is very

small and a global representation of $[\overline{V}]$ has not been available but indirect verification that the term plays a small role in the budgets of momentum and heat has been obtained from the fact that independent estimates of the left hand side of the equation are about the same size as the eddy terms.

On the basis of these meteorological studies it would have been reasonable to anticipate that over a large fraction of the northern hemisphere transient eddies would play a predominant role in the transport of trace substances in the 1000-200 mb layer with standing eddies and mean motions playing secondary roles.

3. Observed Horizontal Motions

a) Zonal motions

Values for the troposphere are available from the work of Buch (1954) for the 1000-100 mb region for the year 1950. As is well known they show a strong westerly jet in middle latitudes in the upper troposphere that moves to the south in winter and intensifies. There are easterlies at low latitudes in both seasons with maximum velocities in the winter. The stratospheric values shown in Fig. 1 are from Murakami

(1962) for 1957 and Dickinson (unpublished) for 1958. These are based on observations from over 220 stations; details of the analysis procedures are presented by the authors elsewhere. The results refer to average values over the three month periods shown. The jet in the upper troposphere extends into the 100-50 mb layer throughout the year. There is a very large increase in speed in the polar stratosphere between the summer and early winter periods as the polar night jet develops. The 10 mb information is not yet adequate to include it in this climatological approach although much analysis has now been performed at that level. At higher levels the zonal components over North America are becoming evident from the Meteorological Rocket Network data. (Webb et al, 1961). The 1960-61 data is shown in Table 1. These summaries have been prepared from the preliminary data published in report form by the U. S. Army Signal Missile Support Agency at White Sands, New Mexico. The data are presented by station as the limited sample does not warrant construction of a meridional cross-section. In winter there are westerly winds increasing with height up to at least 50 or 60 km with mean speeds reaching 60 m sec^{-1}, much greater than those in the tropospheric jet. In summer the high altitude jet is easterly with speeds somewhat smaller than in the winter. Up to 100 observations are included in the summer means and up to 50 in the winter. Some pilot calculations of the eddy flux of momentum in this region by the author (which are being published elsewhere) show a northward flux at the middle

latitude stations. It is likely therefore that this upper jet is maintained by eddy processes in a similar fashion to the jet in the upper troposphere.

By and large the data on the zonal winds suggest that any trace substance introduced into the atmosphere will be distributed fairly rapidly around the globe. The region of slowest zonal circulation, where this does not apply, is the polar stratosphere above 100 mb in the summer season.

b) Meridional motions.

From the point of view of constructing an hemisphere budget for a particular trace substance the meridional motions are of greatest interest. An attempt will be made to divide them into three components appropriate to the three terms in Equation (3).

i) Mean motions.

Much effort has gone into attempts to describe satisfactorily the mean meridional motions in the atmosphere. Buch (1954), working with the northern hemisphere data for 1950 divided into winter and summer seasons, found values of $[\overline{V}]$ that were almost all less than 1 m sec^{-1} and which when combined gave a rather patchy meridional cross-section. Tucker (1959) presented considerable evidence, based upon observations in the region $160^{\circ}W - 0^{\circ} - 40^{\circ}E$, $15-75^{\circ}N$ and 1000-200 mb, over a period of several years, of mean meridional motions with a direct

cell at low latitudes and an indirect cell in middle latitudes. The direct cell was quite strong in the summer and moved further to the north then than in the winter. The northward moving arm of the cell showed velocities up to 1 m sec^{-1} and the southward moving arm up to 2 m sec^{-1}. The main drawback with this work was the limitation in longitude . Recently a set of meridional cross-sections of mean motions and their standard deviations has been presented by Crutcher (1961) for every 10° of longtitude for the 850-100mb layer. Several years data divided into three month seasons has been processed for these cross-sections but there is the drawback that in some regions of poor data winds were actually scaled from pressure charts. The latitudinal average of a mean meridional geostrophic wind should of course be zero. Nevertheless in view of the scarcity of data concerning mean meridional circulations it seemed worthwhile to construct average cross-sections from this work and this was done giving equal weight to each 10° of longtitude. The resultant mean motions showed a three cell structure with a narrow belt (20° in extent) of southward motion (up to 0.5 m sec^{-1}) in summer in middle latitudes which broadened to about 40° in latitudinal extent in winter. The direct cell south of 25° N attained northward velocities of over 2 m sec^{-1} in the winter. The general pattern was that predicted theoretically by

Eady (1950), although much more cannot be said until the work is extended to the surface and attempts are made to satisfy the mass conservation relationship. The MIT work for the stratosphere is shown in Fig. 2; the 1957 data were initially analysed by Murakami (1962) but have been recently modified, while the 1958 data were analysed by Oort and will be published and discussed in detail elsewhere. In the summer there appears to be a reflection of the three cell structure noted for the region below 100 mb with northward motion at both low and high latitudes. Velocities are very small being less than 25 cm sec^{-1} over most of the region in the summer. The values shown are averages of independent analyses made for 0000Z and 1200Z; the differences between these analyses give an estimate of the probable error in the final figures which turns out to be about 15 cm sec^{-1} when all the $[\overline{V}]$ data are taken into account. In the October - December period velocities have increased in most of the region north of 40°N and again southward motion is indicated in middle latitudes with northward motion at high latitudes and essentially no motion south of 30°. In the January - March 1958 period the region of southward motion has broadened and extends to the equator at 50 and 30 mb. Speeds in the majority of this region are 25 cm sec^{-1} whereas in the polar region there is an

extensive area in both winter periods with speeds greater than 50 cm sec^{-1}. In April-June 1958 all the speeds are lower than in the winter with maxima of 50 cm sec^{-1}.

If these mean motions are representative then trace substances introduced into the tropical stratosphere would only drift northwards in summer and then would not reach more than about 30^{o}N. In the winter and spring, when some of the atmospheric models proposed from trace substance observations required a northward drift into middle latitudes, these wind data suggest that the drift is in the opposite direction. Mean meridional motions are not quoted from the Rocket Network data as the sample is too small and limited in longitude.

ii) Standing eddies.

The only method of obtaining an objective assessment of the magnitude of standing eddy velocity components appears to be from consideration of the spacial variance of the mean meridional velocity around a latitude circle. If there are systematic variations of the trace substance concentration with latitude then this meandering may provide transport in the north-south direction. Some substances like ozone might be expected to show such systematic variations due to the association between meandering and the vertical motion. The spacial variances are only presently available for the stratosphere

although estimates for the 850-100mb region are being made from Crutcher's work. Standard deviation values for the first 12 months of the IGY are shown in Fig. 3; they are generally five or more times greater than $[\overline{V}]$ except in the tropical stratosphere. In July-September 1957 the dominant feature of the pattern is the maximum associated with the meandering of the jet in the upper troposphere. In October - December the 50 and 30 mb values are higher by a factor of three to five in association with the polar stratospheric jet. In January-March 1958 even larger values, up to 8 m sec^{-1} at 50 mb, are associated with the polar jet whose effects are at a lower altitude than in the previous period. By April-June 1958 the pattern appears to reflect the effects of jets in the troposphere or lower stratosphere and meandering is again small at 30 mb. The average difference between the 0000Z and 1200Z analyses was about 20 cm sec^{-1}.

(iii) Transient eddies.

The best estimate of this term in equation (3) may perhaps be made from the time variances which give a measure of the north-south exchange of air across latitude circles. At the Symposium Crutcher's values were discussed for the 850-100 mb region for four seasons. The maximum variance is in the 200-300 mb layer and is associated with the jet stream. Values are larger in winter

than in the summer and the maximum moves to the north with the jet in summer as indeed would be expected since the eddy fluctuations have been shown to play a dominant role in the maintenance of the jet. The smallest values are in the lower troposphere in the tropics. Time standard deviations for the stratosphere for 1957 are shown in Fig. 4 (these are preliminary data from Murakami, 1962). The general pattern is similar to that of the space values with the magnitudes somewhat higher, particularly in the tropical stratosphere. Analysis of the 1958 charts has not been completed. For the present problem the differences between the spacial and time variances presumably means that the transient eddies are more important than the standing eddies but there is not an order of magnitude in the differences. Murakami points out that the northward transport of momentum by the standing eddies is comparable with that by the transient eddies in the northern hemisphere.

Very little information is available at higher levels. Time standard deviations calculated from the Meteorological Rocket Network data are shown in Table 2. There is a general increase with height which would be expected if the higher jet is maintained

by quasi-horizontal eddy transports.) Many of the data were obtained with the chaff technique which tends to give an instrumental variance that increases with decreasing altitude.

A brief summary of this section follows. Mean meridional motions exist in both troposphere and stratosphere but are much smaller in magnitude than the eddy motions; the resultant mean motion transport in the middle latitude stratosphere is southward. Lateral exchange is a maximum in the region of the jet stream, concomitant with the tropopause "break" and region of maximum baroclinicity, throughout the year. Another intense region of north-south exchange occurs in connection with the polar night jet stream. The regions of weakest lateral exchange are: the lower tropical troposphere and the polar cap region at and above 30 mb in summer.

4. Vertical Motions

There are no direct measurements of vertical motion in the atmosphere. Indirect estimates can be made either from the horizontal winds and the continuity equation or from the adiabatic assumption. In the former it is necessary to know the winds from the surface upwards; errors tend to accumulate so that values deduced for the higher altitudes are subject to distortion. In the latter the local change of temperature is assumed to be due entirely to the effects of advection and vertical motion; diabatic effects are ignored. While it may be reasonable to take

this approach for a value representative of a short time period and when the vertical velocity is large, the procedure cannot be used specifically to calculate long term average vertical velocities where radiative effects are known to contribute to the temperature changes. It is only recently that a beginning has been made on the calculation of vertical flux processes in the atmosphere from these imperfect vertical velocities. Jensen (1960) and Barnes (1962 unpublished) have worked on these processes in the regions 1000-50 and 100-30 mb respectively and have calculated vertical velocities on a daily basis and then computed climatological averages. There is not yet a clear cut explanation of the relative roles of the various types of motion in the vertical transport problem. The available vertical motion data will be briefly reviewed below.

(i) Mean motions.

Tucker (1959) has used the continuity approach to calculate the mean vertical velocities from the mean meridional winds. He finds a downward motion centered at about 25°N in winter with an average value of about 0.5 cm sec^{-1}, and a rising motion of about 0.2 cm sec^{-1} at about 42°N. In the summer both cells move towards the north maintaining about the same intensity. The motions would tend to produce a belt of high concentration of a trace substance (having its source in the stratosphere) at about 30°N in winter and 35°N in summer. But note that the speeds are rather low -- it

would take a small particle over 20 days to pass from the tropopause to the surface. Murakami (1960), by the same approach, has found a similar structure for the mean vertical motions over the hemisphere.

Barnes has investigated the mean motions in the stratosphere from the IGY data using 220 stations and the adiabatic approximation. Daily values of vertical velocity were calculated by machine methods for each station and plotted on hemispheric maps. Isolines were drawn and values at every 10° of longitude and 5° of latitude were tabulated and averaged.

Three month seasonal means were calculated for 40 mb and 75 mb and will be published elsewhere by Barnes. In the July - September period the resultant velocities at 40 mb are practically zero whereas at 75 mb there are rising motions to the north of 40°N with sinking to the south. In the October-December period the rising motions extend up to 40 mb north of 45°N with essentially zero to the south; at 75 mb the rising motions north of 40° N are somewhat larger than in the previous period. The general order of magnitude of these mean motions is 0.05 cm sec^{-1}.

(ii) Standing eddies.

The analyses are not yet complete.

(iii) Transient eddies

As in the case of horizontal motions values of the variance of the vertical velocity will be considered as a representation of the degree of vertical mixing of the atmosphere. Jensen (1960) has presented values of the adiabatic vertical velocity and its standard deviation for 90 stations over the northern hemisphere for January and April 1958. From the data tabulated in Jensen's report the stations were divided into latitude belts and the standard deviations averaged for each belt objectively. The results appear in Tables 3 and 4. The cross-sections are not particularly smooth because of the small data sample used. In January maximum values are in the upper troposphere, in the vicinity of the jet. Minimum values are in the 100-50 mb layer and also in the tropical troposphere. In April there is the same general pattern but with somewhat lower values with a marked decrease in the polar stratosphere. The stratosphere standard deviations for the IGY period appear in Table 5. South of 50°N the maximum values, which are at most twice the minimum values at the same latitude occur in the January-March 1958 period. Farther north at 60° and 70° the maxima occur in the October-December period.

The turbulent velocities in the troposphere are seen to be much greater than the mean vertical motions. In the stratosphere too the eddy components are greater than the adiabatic means. It would be reasonable to expect that trace substances would be transported in the vertical more by eddy motions than by mean motions.

Much of the vertical velocity variance may be closely related to the horizontal variance insofar as it may be the vertical component of quasi-horizontal transient eddy motions which is being recorded.

5. The Inadequate Knowledge About Turbulent Diffusion

If a trace substance is introduced into the atmosphere at a particular point it will diffuse from the point depending on conditions of wind and stability. It is ofter customary to express the diffusive properties of the medium in terms of an eddy diffusion coefficient. The problem is to decide how to calculate that the coefficient from the observed meteorological variables. For the purpose of discussion at the Symposium diffusion coefficients were represented as the time variances multiplied by a constant (which was arrived at by arranging for the horizontal diffusion coefficient over a large area near the jet to be 10^{10} cm^2 sec^{-1}). The patterns then simply reflected the variance patterns. For January 1958 the calculated vertical eddy diffusion in the stratosphere was of order 10^3 near the equator and about 10^4 near the pole. The smallest horizontal coefficients on this hypothetical scheme were $10^{8.6}$ cm^2 sec^{-1} in the tropical troposphere and $10^{8.8}$ cm^2 sec^{-1} in the lower tropical stratosphere. The vertical eddy coefficients in the troposphere were about 10^5 cm^2 sec^{-1} in the vicinity of the

jet and only $10^{3.5}$ cm^2 sec^{-1} in the lower tropical troposphere. The latter region has the smallest values of both parameters except for the equatorial stratosphere. The vertical coefficients are undoubtedly related to the static stability; the stratospheric temperature structure is shown in Fig. 5. Certainly the smallest variances accompany the large static stability in the equatorial regions. But note that small values also occur in the tropical troposphere. Factors other than the lapse rate must govern the fluctuations. It is not at present clear that the diffusion coefficient approach can provide a complete basis for a model as any possible thermodynamic or dynamic constraints such as a tendency towards conservation of potential temperature of potential vorticity cannot yet be included.

6. Application to Trace Substances

Given the observed meteorological structure reviewed in previous sections it is instructive to examine some of the features of the global distributions of trace substances to see if they can be satisfactorily explained.

Consider first the case of the long-lived fission product radioactive substances. There are three outstanding features

of the surface air concentration of these substances (a) they exhibit a maximum in middle latitudes in both hemispheres regardless of the injection latitude (b) the maxima move north and south with the sun and (c) maximum concentrations occur each year in the spring. These three features have been discussed in detail elsewhere (Newell, 1961). It has been suggested that the reason for (a) is the quasi-horizontal stratospheric-tropospheric exchange which occurs most readily in the vicinity of the jet stream (using the word now in its broadest sense). These long lived fission products can be considered to have their source in the stratosphere as they are efficiently removed from the troposphere by precipitation after a mean life of about 30 days. This exchange of air has been discussed in detail by several authors (for example, Danielsen, 1960; Staley, 1960). There is also certain experimental evidence to support the meteorological conclusions such as the ozone soundings of Brewer (1960) (which indicate ozone-poor layers in the lower middle-latitude stratosphere which may have originated in the tropical upper troposphere), and Ney and Kroening (1962) (which indicate ozone-rich layers in the upper troposphere that may have entered from the stratosphere) and the negative correlation between ozone and water vapor concentrations observed by Roach (1961, unpublished) in the vicinity of

the jet. Even if these quasi-horizontal exchanges are the main mass exchange mechanism it is still possible that direct exchange occurs at high latitudes, particularly in the winter season. Some indirect evidence of this has been presented elsewhere (Newell, 1960).

Suppose that for purposes of further discussion it is accepted that the majority of the tropospheric-stratospheric mass exchange occurs in the vicinity of the middle-latitude baroclinic zone or jet stream then feature (a) would be explained and as the jet moves to the north in the summer so also would feature (b). If the exchange is more or less along isentropic surfaces which pass from the stratosphere into the troposphere then the spring maximum could be due either to a greater mass exchange along these surfaces during the spring or to a larger gradient of concentration such as may be brought about by larger concentrations in the polar and middle-latitude lower stratosphere during the late winter than during any other season. The horizontal variance data in the vicinity of the jet do not have a very large seasonal change. In the summer of 1959 the tropospheric concentrations of gross β - activity declined at a rate which was almost equal to the washout rate, indicating that there was a very small influx of contaminated air into the troposphere during the period. It

would therefore seem that the origin of the spring maximum must be sought in the second circumstance suggested above. The problem than becomes to explain an additional communication between the source regions and the middle latitude lower stratosphere (from whence the debris can pass into the troposphere) during the late winter. As has been noted elsewhere the debris from both major source regions, the Pacific tests held at low latitudes and the Russian Arctic tests, exhibits a similar spring maximum. Thus the mean motions cannot be the main contributor to this effect. Friend, Feely and Krey have discussed at this Symposium the tungsten data which demonstrate that the centre of gravity of the tungsten cloud did not move significantly in latitude under the action of a mean motion but that the principal method of transport polewards appeared to be by eddy diffusion. The author (Newell, 1961) has elsewhere made similar claims for ozone. Thus the spring maximum might be explained by large eddy diffusion effects in late winter that transfer all the trace substances down their concentration gradients and into the region of the lower stratosphere favourable for exit into the troposphere. The variance data presented for the stratosphere do have maximum values in the late winter period in both vertical and horizontal components in middle latitude and polar regions. Such mixing would also transfer rhodium-102 down

from its source above 30 mb as reported by Kalkstein. Unfortunately our data on the time variance of the meridional wind were not completed for 1958 in time for the Symposium but the general parallel between the time and space variances for 1957 probably is a feature of northern hemisphere flow patterns. Crutcher's time standard deviations at 100 mb appear to be a factor of two larger in the December-February period than in the summer in the polar stratosphere. It would therefore seem that extra lateral and vertical exchange occurs during the spring period which is responsible for mixing debris into the region from which it has access into the troposphere; the mixing occurs down the gradient of a particular trace substance, whether it be Pacific tungsten, Novaya Zemlya strontium or ozone.

7. A Note on Possible Motions Involved

The next two diagrams shown at the Symposium were meridional cross-sections of tungsten-185 (published by Friend et al, 1961) superimposed on the mean isentropic surfaces. As these diagrams have been published elsewhere (Newell, 1962) for other purposes they will not be reproduced here. It appears that tungsten spread polewards

and downwards from the equatorial lower stratosphere following the general slope of the isentropic surfaces. But clearly the debris cannot be confined to isentropic surfaces otherwise it would never leave the stratosphere. In fact if the concentration isolines of strontium or ozone are examined then they appear to slope downwards between low latitudes and 50-60°N at a greater slope than the mean isentropes. Some possible reasons for the difference in slope were discussed elsewhere (Newell, 1961).

In the middle latitude troposphere differential heating produces a slope in the isentropic surfaces which can be viewed as the production of available potential energy. Disturbances arise to transport the heat northward and they take advantage of this available potential energy and in fact run on it by a process in which warm air rises and cold air sinks. These same disturbances are responsible for the transport of angular momentum northwards. In order to release potential energy the air motions must be across the isentropes. Eady (1950) and Green (1960) have shown that for optimum efficiency the slope of the motions in the eddies should be about 1/2 of the slope of the mean isentropes. These eddies will then constantly be tending to flatten the isentropes while

the differential heating tends to steepen them. There is an essentially similar situation during the polar night vortex development. Information about the eddies in the stratosphere is meagre in this vein. Martin and Brewer (1959) found from trajectory analysis in the lower stratosphere over Europe that air parcels which had moved from the south contained more ozone and were warmer than those which had moved from the north. This is in agreement with the tungsten data. Molla and Loisel (1962) correlated the adiabatic vertical velocities used by Jensen (1960) with the northward component of the wind for the months of January and April 1958. In the 100-50 mb layer they found a negative correlation south of 50°N in January and a positive correlation to the north which suggests that the general motions tend to follow the isentropes. In April the negative correlations are smaller south of 50°N and essentially zero north of 60°N, again in approximate agreement with the isentropes. These correlations can explain the observations of a countergradient heat flux in the lower stratosphere noted by White (1954), Peixoto (1960) and Murakami (1962) if the additional hypothesis is made that the motions are steeper than the isentropes (that is tilt downwards towards the north more than the isentropes) and in this case the

motions will be tending to convert kinetic energy into potential energy. Preliminary examinations of limited samples of stratospheric data by White and Nolan (1960) and Jensen indicate conversions in this sense by transient eddies. Sufficient data is not yet available to attempt explanations for the regions north of 60^{o} N.

8. The Application of the Transport Equation to Ozone

As part of our calculation procedure we have correlated the ozone amounts reported during the IGY with the wind components, temperatures, heights and vertical velocities in the stratospheric region. Some of the correlations have been reported elsewhere (Newell, 1961) where it was pointed out that the total amounts of ozone could be used as an indicator of the amount of ozone in the lower stratosphere. The results for the eddy flux show a northward flux in middle latitudes which has a maximum in the spring season. The results for the four terms in equation (3), each evaluated objectively, are shown in Table 6. The left hand side of the equation was calculated from the changes in the total amount of ozone in the polar cap. The downward flux in the troposphere was evaluated from the work of Regener (1957). It provides an additional amount of ozone which may be passing through the entire system; there is currently no information available about its variation with latitude and season. The eddy terms appear to be sufficient in the spring period

to account for the observed build-up of ozone even allowing for a continual drain. It is hypothesized that ozone is moved northwards and downwards from the major source region in the low-latitude stratosphere by eddy mixing, and is then mixed in the polar and middle latitude stratosphere until a good fraction reaches the region in the lower stratosphere from which it can pass into the troposphere. If the northward eddy fluxes are considered to be cut off then the polar stratosphere below about 25 km could be completely emptied of ozone in four months by flux into the troposphere. The mean life of other trace substances in the same portion of the lower stratosphere might be expected to be similar. In conclusion it must be emphasized that in order to extend this budget procedure to other trace substances it is imperative to obtain observations over long time periods at a number of stations over the globe.

Acknowledgements

As mentioned in the text I am grateful to Professor Starr and my colleagues on the Planetary Circulations Project for use of their unpublished data and for many discussions. Mrs. Barbara Goodwin and Mrs. Dorothy Berry have again looked after the computational work. It is also a pleasure to thank Professor Sheppard at Imperial College, London, for hospitality while part of this paper was written particularly for his discussions concerning the topics of Section 7.

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Table 1 METEOROLOGICAL ROCKET NETWORK. Mean zonal winds $\bar{u}$ meters sec^{-1}

(km)	Height (ft)	1 58.8N 94.3W	2 38.0N 116.5W	3 37.8N 75.5W	4 34.1N 119.1W	5 32.3N 106.5W	6 32.9N 106.1W	7 30.5N 86.5W	8 28.2N 80.6W
	Winter 1960-61								
61.0	200,000	31.9	27.6	51.6	54.5	55.2	58.3	43.3	57.8
54.9	180,000	56.9	25.0	42.3	46.5	53.4	31.2	51.4	59.1
48.8	160,000	48.7	12.9	39.7	40.6	49.4	31.9	56.3	43.0
42.7	140,000	23.9	35.3	31.6	33.8	41.0	27.8	48.3	35.3
36.6	120,000	17.9	13.8	22.6	22.7	24.5	17.0	37.0	27.0
30.5	100,000	8.9	4.9	12.2	11.4	8.5	4.2	16.5	11.3
24.4	80,000	8.7	-0.1	3.8	2.8	4.4	9.0	2.6	5.6
18.3	60,000	7.4	5.5	12.1	10.0	11.1	10.4		19.2
	Summer 1960-61								
61.0	200,000			-28.0	-51.9	-33.1			-20.5
54.9	180,000			-25.3	-31.4	-34.2			-31.1
48.8	160,000			-22.1	-29.7	-28.3	-12.7		-34.5
42.7	140,000	-19.2	-27.5	-19.4	-22.8	-21.5	-13.1		-27.8
36.6	120,000	-13.6	-19.3	-14.9	-14.4	-16.7	-10.4		-20.0
30.5	100,000	-10.8	-15.0	-12.4	-13.3	-14.9	-7.9		-18.0
24.4	80,000	-4.1	-11.9	-8.1	-11.4	-13.3			-16.1
18.3	60,000	+1.0		-0.8	-3.1	-3.4			+0.5

1. Ft Churchill, Canada
2. Tonopah Range, Nev.
3. Wallops Is., Va.
4. Pt. Mugu., California
5. White Sands, N.M.
6. Holloman, N.M.
7. Eglin Field, Florida
8. Cape Canaveral, Florida

Table 2 METEOROLOGICAL ROCKET NETWORK. Standard deviation of mean meridional wind $\sigma(V)$ metres sec^{-1}.
Number of cases in parentheses

(km)	Height (ft)	1 58.8N 94.3W	2 38.0N 116.5W	3 37.8N 75.5W	4 34.1N 119.1W	5 32.3N 106.5W	6 32.9N 106.1W	7 30.5N 86.5W	8 28.2N 80.6W
Winter 1960-61									
61.0	200,000			9.9 (14)		12.6 (32)		8.2 (11)	
54.9	180,000			11.1 (29)	13.5 (26)	12.4 (40)		8.3 (12)	11.4 (11)
48.8	160,000			11.4 (38)	8.8 (39)	12.3 (40)	13.0 (12)	9.9 (14)	14.9 (17)
42.7	140,000	9.6 (15)	8.8 (15)	11.3 (47)	8.8 (44)	8.2 (43)	14.6 (16)	12.0 (15)	10.5 (16)
36.6	120,000	14.5 (24)	6.5 (36)	9.0 (51)	6.2 (55)	5.4 (56)	7.9 (19)	5.4 (14)	7.6 (21)
30.5	100,000	12.3 (30)	5.6 (36)	5.5 (57)	5.5 (63)	5.1 (59)	4.3 (19)	3.1 (11)	4.4 (21)
24.4	80,000	8.8 (33)	3.5 (30)	3.8 (42)	4.0 (60)	3.8 (62)			2.7 (12)
18.3	60,000	5.3 (24)	4.8 (16)	7.0 (31)	4.6 (49)	4.7 (45)			
Summer 1960-61									
61.0	200,000					8.3 (29)			16.1 (11)
54.9	180,000			4.8 (11)	6.0 (28)	9.4 (52)			8.0 (32)
48.8	160,000			6.0 (28)	4.9 (44)	5.2 (67)			8.2 (43)
42.7	140,000			4.1 (29)	4.1 (53)	5.8 (77)			6.0 (45)
36.6	120,000		3.2 (14)	2.8 (31)	3.2 (54)	4.4 (99)			5.4 (42)
30.5	100,000	5.6 (20)	2.8 (14)	3.3 (31)	3.0 (56)	4.0 (104)			3.5 (41)
24.4	80,000	6.1 (25)	2.2 (14)	2.1 (27)	1.9 (49)	2.5 (112)			2.2 (15)
18.3	60,000	4.6 (21)		3.7 (26)	3.1 (37)	3.1 (97)			

1. Ft. Churchill, Canada
2. Tonopah Range, Nev.
3. Wallops Is., Va.
4. Pt. Mugu., California
5. White Sands, N.M.
6. Holloman, N.M.
7. Eglin Field, Florida
8. Cape Canaveral, Florida

Table 3 Standard deviation of Vertical Velocity $\sigma(w)$ cm sec^{-1}
January 1958

Pressure layer (mb)	Latitude belt (degrees) 0-12	13-17	18-22	23-27	28-32	33-37	38-42	43-47	48-52	53-57	58-62	62-72	73-83
100-50	0.3	0.3	0.4	0.7	0.6	0.6	0.5	0.7	0.7	0.8	0.6	0.7	1.0
200-100	1.0	1.1	1.3	1.9	1.5	0.9	0.8	0.8	0.9	1.1	0.8	0.7	1.7
300-200	2.7	2.3	3.4	10.8	9.6	4.8	3.6	2.6	2.8	2.5	3.6	1.0	1.3
500-300	1.2	1.3	1.9	3.3	5.1	5.5	3.6	3.2	3.7	4.2	3.9	2.6	2.3
700-500	[illegible]	0.9	1.2	2.1	3.4	2.8	2.3	2.7	2.7	3.2	2.6	2.5	2.2
850-700	0.5	[illegible].5	0.8	1.9	1.9	1.9	2.3	2.0	2.2	2.5	2.0	2.0	2.1
1000-850	0.9	0.[illegible]	1.9	2.0	2.8	4.0	2.1	2.5	2.7	3.0	1.2	1.1	2.4

Table 4 Standard deviation of Vertical Velocity σ (w) cm sec^{-1}

April - 1958

Pressure layer (mb)	Latitude belt (degrees) 0-12	13-17	18-22	23-27	28-32	33-37	38-42	43-47	48-52	53-57	58-62	62-72	73-83
100-50	0.3	0.3	0.4	0.7	0.8	0.7	0.4	0.5	0.5	0.6	0.3	0.6	0.3
200-100	0.9	1.0	1.3	1.4	1.3	0.8	0.7	0.7	0.6	0.6	0.4	0.5	0.3
300-200	1.9	2.4	3.3	4.2	5.6	4.7	4.0	3.6	2.3	2.0	1.1	1.5	0.4
500-300	0.8	1.2	1.9	2.4	3.7	3.6	3.9	3.7	3.4	3.7	2.6	2.6	0.9
700-500	0.6	0.7	1.0	1.7	2.3	3.1	3.2	2.3	2.5	2.5	1.6	2.0	1.9
850-700	0.5	0.6	0.7	1.7	2.3	2.8	4.9	2.8	3.3	3.1	1.7	1.3	1.7
1000-850	1.0	1.0	1.6	1.3	2.3	2.3	2.2	3.5	4.1	3.2	3.1	1.5	1.2

Table 5 Time Standard Deviation of Vertical Velocity

$\sigma(W)$ cm sec^{-1}

Pressure layer	Lat.	10°	20°	30°	40°	50°	60°	70°	80°
50-30mb									
July-Sept.	1957	0.19	0.24	0.27	0.33	0.33	0.28	0.25	0.22
Oct-Dec.	1957	0.22	0.22	0.27	0.30	0.46	0.59	0.61	0.38
Jan-Mar.	1958	0.22	0.28	0.31	0.46	0.64	0.57	0.51	0.41
April-June.	1958	0.23	0.27	0.27	0.31	0.34	0.36	0.41	0.34
100-50mb									
July-Sept.	1957	0.26	0.28	0.31	0.37	0.35	0.28	0.27	0.21
Oct-Dec.	1957	0.22	0.24	0.33	0.40	0.48	0.62	0.61	0.38
Jan-Mar.	1958	0.25	0.36	0.44	0.47	0.48	0.50	0.55	0.56
April-June	1958	0.23	0.26	0.29	0.34	0.32	0.32	0.27	0.24

Table 6

OZONE BUDGET

Units are in cm^3 of O_3 at STP sec^{-1} when multiplied by 10^9.

	July-Sept. 1957	Jan-Mar. 1958	July-Sept. 1958
Transport across 50N.			
a) by mean meridional motions	-0.9	-1.9	?
b) by standing eddies	+0.3	+4.0	-1.0
c) by transient eddies	+2.9	+10.4	+2.0
From content change	-5.2	+9.0	-5.2
Transport across 40N.			
a) by mean meridional motions	-2.3	-3.8	?
b) by standing eddies	+0.4	+5.0	-1.2
c) by transient eddies	+0.4	+12.6	+0.5
From content change	-7.4	+11.6	-7.4

Tropospheric downward flux ~ 6.

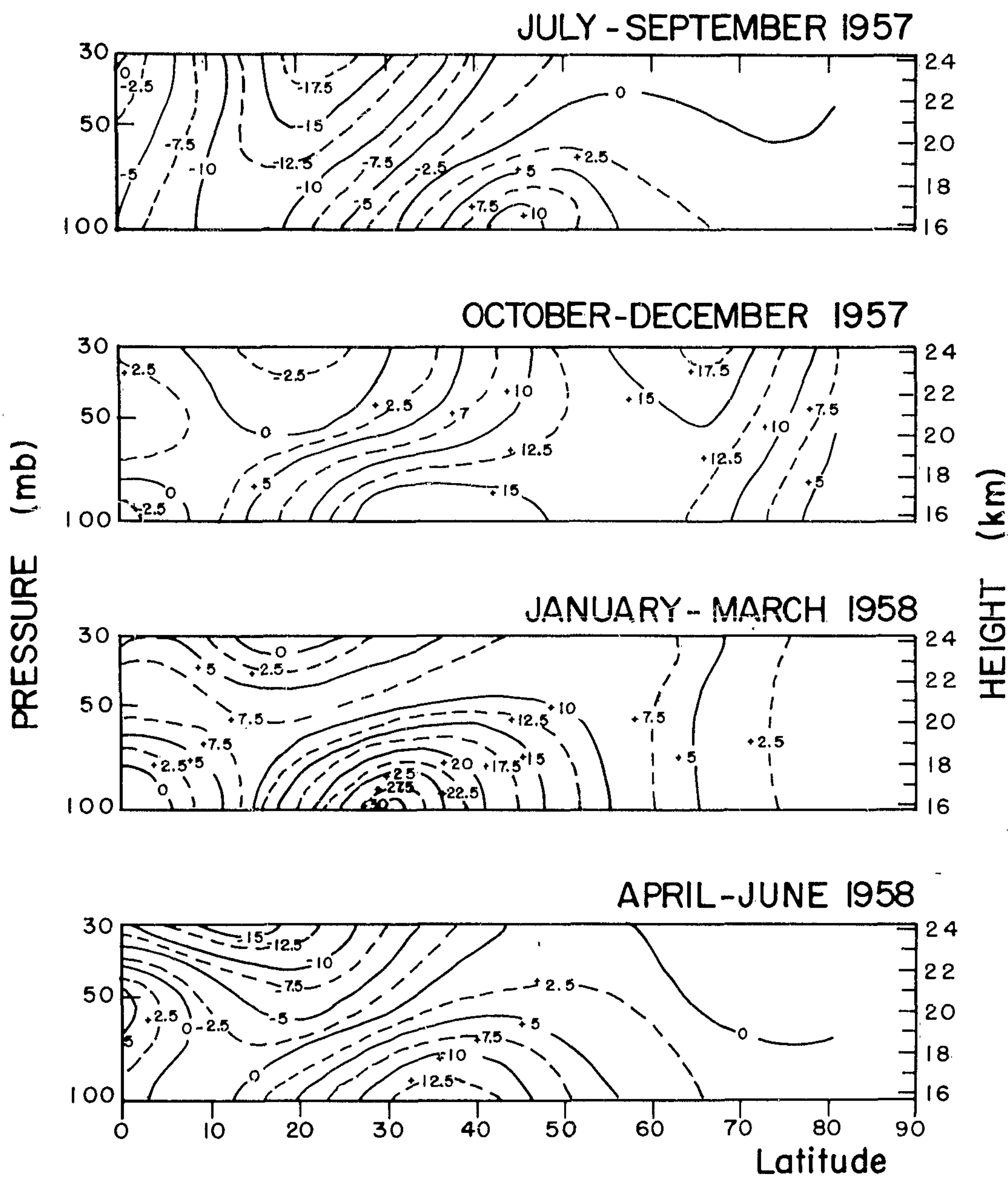

Figure 1. Mean zonal velocity. $[\bar{u}]$. metres sec^{-1}

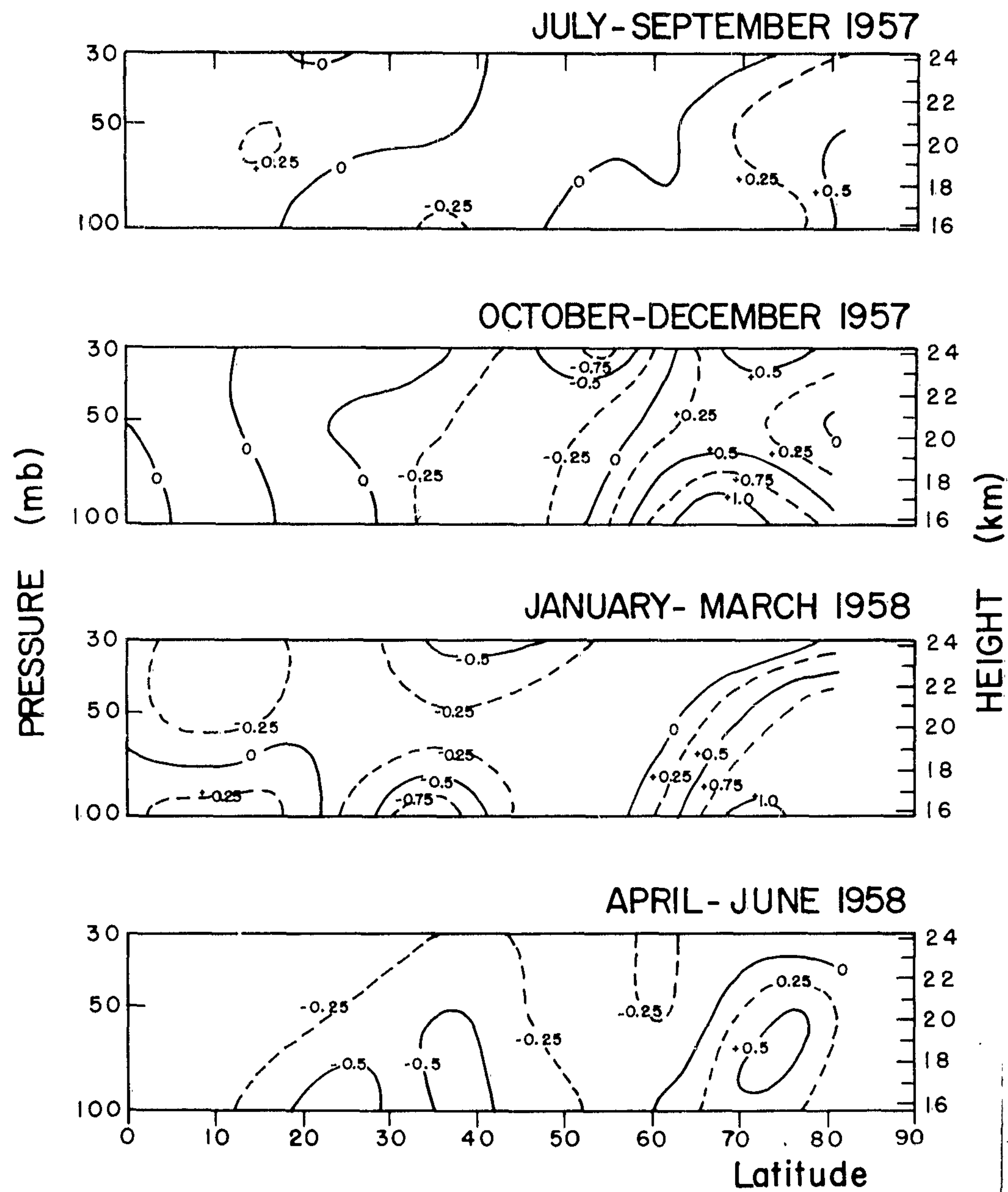

Figure 2. Mean meridional velocity. $[\overline{V}]$ metres sec^{-1}

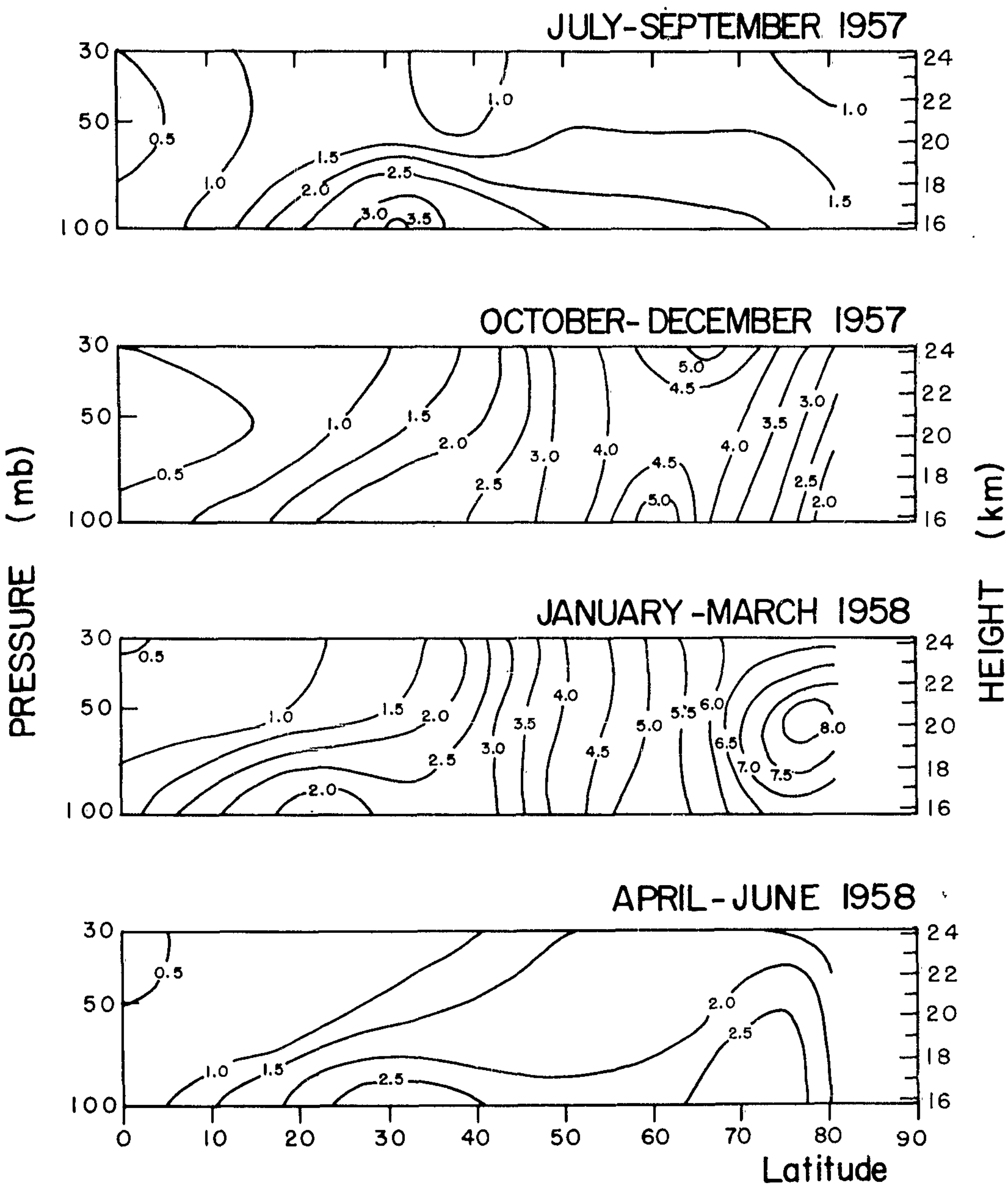

Figure 3. Spacial standard deviation of standing eddies component of V . $\sqrt{[\overline{V^{*2}}]}$. metres sec^{-1}

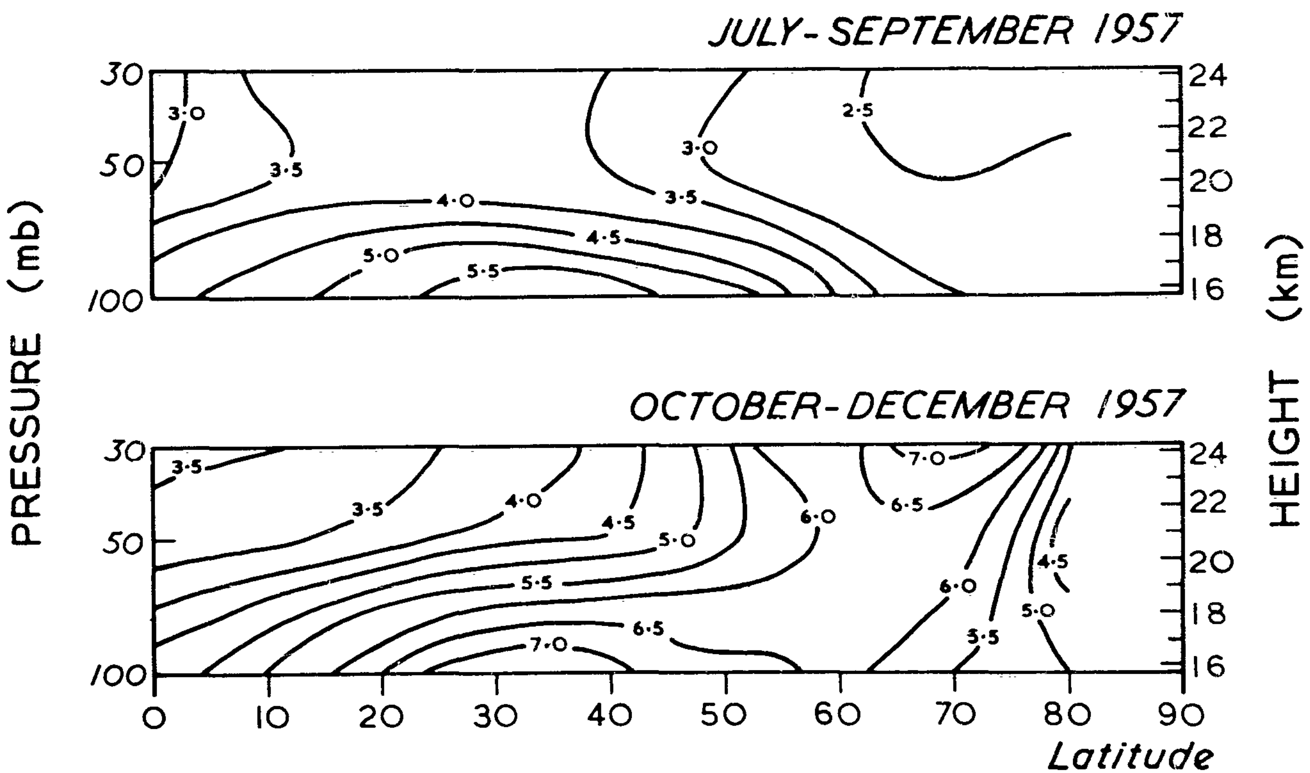

Figure 4. Zonal mean of time standard deviation of transient eddies component of V $\left[\sqrt{\overline{V'^2}}\right]$ metres sec^{-1}

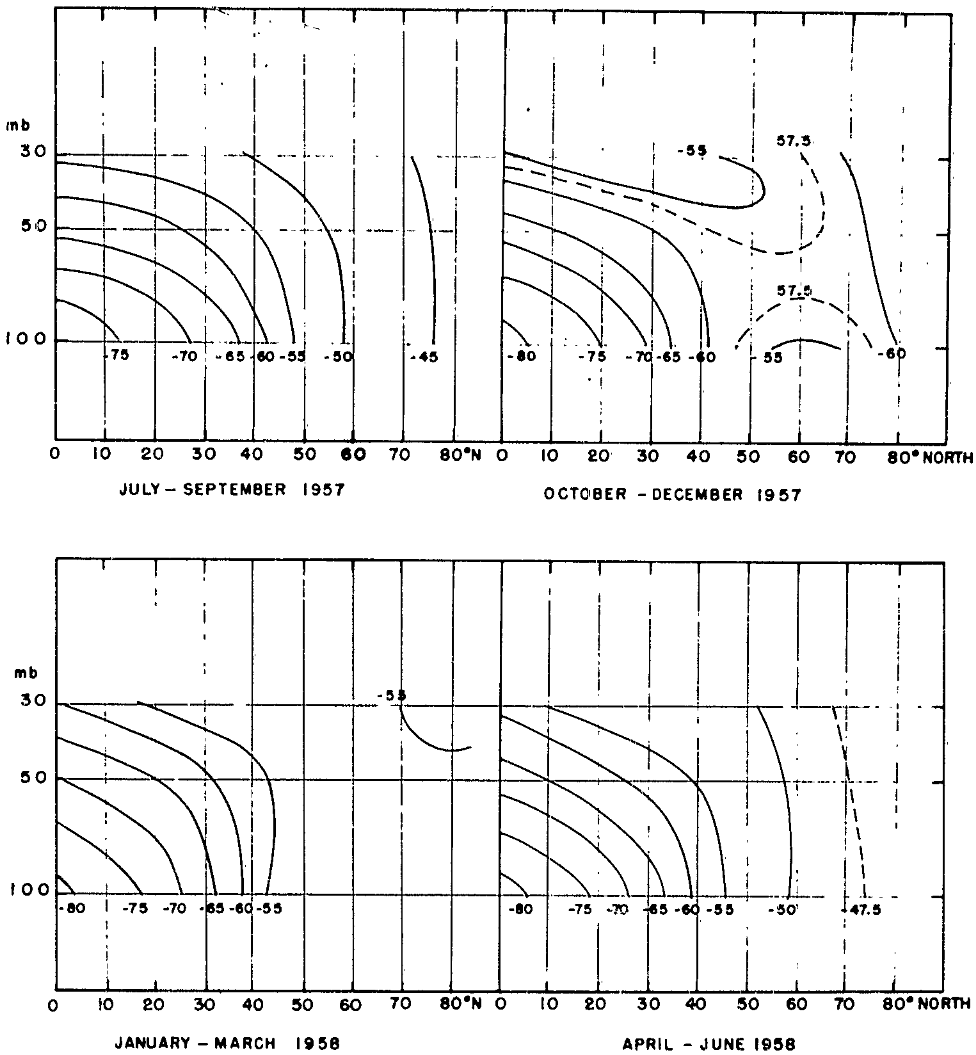

Figure 5. Zonal mean temperature. [T] °C.

APPENDIX

REPRINTED FROM
MONTHLY WEATHER REVIEW

JANUARY 1963

ANALYSIS OF THE U AND V FIELDS IN THE VICINITY OF THE POLES[1]

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[Manuscript received August 7, 1962; revised October 18, 1962]

ABSTRACT

The fields of the eastward (U) and northward (V) components of the wind are investigated mathematically. Maps whose point values are the products of the point values of two maps are also investigated. Rules are given to aid in the analysis of these types of maps.

1. THE PROBLEM

The use of polar coordinate charts means that the maps of U and V, the eastward and northward components of wind, have mathematical singularities at the poles. The existence of these singularities has been either completely ignored or circumvented by some device such as, for example, not analyzing near the poles. A strict mathematical analysis of the U and V fields in the vicinity of one of these poles reveals information which has proved to be of great assistance to the analyst. Application of our rules considerably improved the analysis of a series of Northern Hemisphere maps (Barnes [1]) in the region north of 75° N.

Even though the following mathematical development is around the North Pole, it is applicable to the pole of any polar coordinate system such as, for example, those frequently employed in special studies of hurricanes and extratropical storms where the radial, V, and tangential, U, velocity components are used.

One of the ways of representing the horizontal wind field is to use the south to north component, V, and the west to east component, U. This representation is unique except at the north and south poles. A closer inspection reveals that, under the U, V decomposition, one is mapping a cylinder onto a spherical surface (e.g., 100-mb. surface). This mapping of a cylinder onto a coaxial spherical surface is unique (one-to-one) except at the poles. The top and bottom lines of the cylinder are mapped onto the poles so the mapping is not one-to-one at these points. Thus, special consideration must be given to quantities containing U or V in the neighborhoods of the north pole and the south pole.

2. THE MATHEMATICAL ANALYSIS

Let us consider the wind over a small neighborhood centered at the north pole. If we take this neighborhood small enough, then the quantities such as the horizontal wind velocity, temperature and height of the pressure surface may be considered as constant. This assumption seems to be better justified in the stratosphere than near the surface of the earth where the temperature and horizontal velocity fields may be discontinuous. The fact that one performs continuous, smooth analysis of the maps using discrete data presumes that the gradients are smooth and the point values are finite.

In a sufficiently small neighborhood of the north pole, the horizontal wind velocity can be considered constant and can be defined in the following manner. First, redefine the longitudes so that λ west longitude becomes $(360°-\lambda)$ east longitude. Then the horizontal wind is uniquely defined by the non-negative quantity C, the wind speed, and λ_c, the meridian along which the wind is blowing toward the north pole. If ϵ is the distance from the pole, then ϵ_M is defined as the radius of the largest circle which fits inside the neighborhood of the pole where the horizontal wind velocity can be considered constant. The map of V is now given by $C\cos(\lambda-\lambda_c)$, $0<\epsilon<\epsilon_M$.

If, in this neighborhood, we approach the pole along the meridians $(\lambda_c+90°)$ and $(\lambda_c-90°)$ we find the value of V to be zero, no matter how close we come to the pole along these lines. A map shall be said to be a *pattern 1 map* if it can be expressed by $C\cos(\lambda-\lambda_c)$ for $\epsilon>0$ in the neighborhood of the pole. Thus, figure 1, showing the map of V in the neighborhood of the north pole assuming a wind speed of 10 kt. is a pattern 1 map.

The map of the U component is given by $C\sin(\lambda-\lambda_c)=C\cos[\lambda-(\lambda_c+90°)]$, so the map of U is also a pattern 1 map. Thus, the U map is just the V map rotated 90° to the east about the pole. This means that for the same time period, the zero isopleths on the maps of U and V should be at right angles to each other in the neighborhood of the pole.

As mentioned before, the U and V components of the horizontal wind are not defined at the pole. Hence, the pole has been excluded from figure 1. If the value at the pole were assigned the zero value, then the zero isotach would be continuous and would not have a kink at

[1] The research reported in this document was carried out while the author was at the Massachusetts Institute of Technology, Cambridge, Mass., and was supported by the Geophysics Research Directorate of the Air Force Cambridge Research Laboratories, Office of Aerospace Research, under Contract No. AF19(604)-5223 and by the U.S. Atomic Energy Commission under Contract No. AT(30-1)-2241.

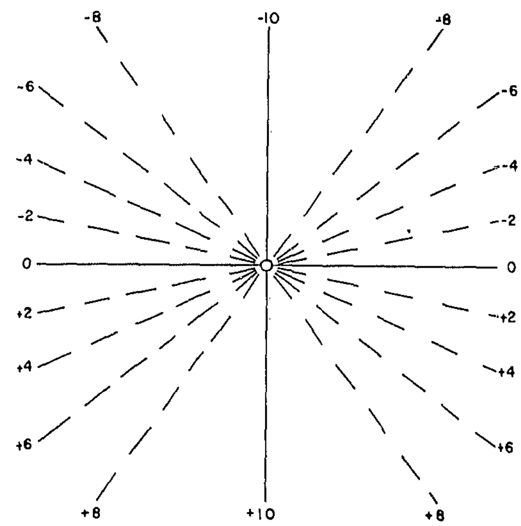

FIGURE 1.—A pattern 1 map, $C \cos(\lambda-\lambda_c)$, for $C=10$ kt.

the pole. If we consider any other contour K $(-C \leq K \leq C)$ and let the pole have the value K, then this isotach would have a kink at the pole. Notice that the equation $U^2+V^2=C^2$ does not hold at the poles since U and V are not uniquely defined at the poles.

In the following discussion we shall be concerned with the neighborhood of the pole and not with the value at the pole itself.

As stated before, a pattern 1 map is represented mathematically by $C \cos(\lambda-\lambda_c)$ for $0<\epsilon<\epsilon_M$, where C and λ_c are constants, but is undefined for $\epsilon=0$.

The sum of two maps is defined as the map of the sum of the values at the individual points. Adding two pattern 1 maps, we obtain, for $0<\epsilon<\epsilon_M$,

$$C_1 \cos(\lambda-\lambda_1)+C_2 \cos(\lambda-\lambda_2)=\cos\lambda[C_1 \cos\lambda_1 + C_2 \cos\lambda_2]+\sin\lambda[C_1 \sin\lambda_1+C_2 \sin\lambda_2].$$

If we now let

$$C_3 \cos\lambda_3=C_1 \cos\lambda_1+C_2 \cos\lambda_2$$

and

$$C_3 \sin\lambda_3=C_1 \sin\lambda_1+C_2 \sin\lambda_2$$

then

$$C_1 \cos(\lambda-\lambda_1)+C_2 \cos(\lambda-\lambda_2)=C_3 \cos\lambda \cos\lambda_3 + C_3 \sin\lambda \sin\lambda_3=C_3 \cos(\lambda-\lambda_3)$$

where

$$C_3=\sqrt{C_1^2+C_2^2+2C_1C_2 \cos(\lambda_2-\lambda_1)}$$

Thus, pattern 1 is conserved under addition.

The negation operation is defined by multiplying all point values of the map by minus one. This is equivalent to replacing λ_c by $(180°+\lambda_c)$ giving $C \cos(\lambda-180°-\lambda_c)$ for V, which is only a change of orientation. Thus, pattern 1 is conserved under negation. It should be noted that the orientation is not unique unless C is positive.

Subtraction of map X from map Y is performed by the subtraction of the point values of X from the point values of Y. This is equivalent to the addition of map Y and the negative of map X, so pattern 1 is conserved under the operation of subtraction.

If the value at every point of a pattern 1 map is divided by a finite non-zero number of quantity Z which is a constant over the map, then the resulting map,

$$\frac{C \cos(\lambda-\lambda_c)}{Z}=\left(\frac{C}{Z}\right)\cos(\lambda-\lambda_c)$$

is a pattern 1 map, and pattern 1 is conserved under division by a finite non-zero constant. If Z is negative, there is also a change of orientation by 180°, but the map remains a pattern 1 map.

Since pattern 1 is conserved under addition and division by a finite non-zero constant, the mean map of a finite number of pattern 1 maps is also a pattern 1 map.

Let us consider the field of the product quantity (XV) in the neighborhood of the pole. If X has the same value throughout the neighborhood, $0<\epsilon<\epsilon_M$, then the map of (XV) will be a pattern 1 map in this neighborhood. Most meteorological variables such as temperature, pressure, height of pressure surfaces, ozone concentration, vertical wind speed, and horizontal wind speed are usually considered as continuous quantities in the free atmosphere. The usual analysis of maps of such variables from a finite number of observations presumes that the fields of these variables do not contain discontinuities. Hence, for most purposes maps of (XV) and (XU), $0<\epsilon<\epsilon_M$, will be pattern 1 maps when X is one of the above meteorological variables.

We should look at two of the unusual cases which appear at times. The first is when X is discontinuous in the neighborhood of the pole, and the second is when X has a zero isopleth in the neighborhood of the pole.

Since the fields of U and V may always be considered as continuous except at the poles, the discontinuities in the fields of (XU) and (XV) may occur only at the poles and where X is discontinuous. (At the point where a zero isopleth of a continuous field X crosses a discontinuity in the field of Y, (XY) is continuous.)

Since the product (XY) is zero when and only when either X or Y is zero, the field of (XY) will have zero isopleths where either X or Y have zero isopleths. (This assumes that X and Y take on only finite values.) Thus, if X is discontinuous or has a zero isopleth through the pole, the maps of (XU) and (XV) will not be pattern 1 maps in the neighborhood of the pole.

The covariance of X and Y is:

$$\text{cov}(X,Y)\equiv\left(\frac{\Sigma(XY)}{N}-\frac{\Sigma X}{N}\cdot\frac{\Sigma Y}{N}\right)$$

Thus, for a finite N greater than 1, if X and Y are pattern 1 maps, then the map of cov (X, Y) is also a pattern 1 map.

As a corollary, we can prove that there are perpendicular zero isopleths on the maps of (UV) in the neighborhood of the pole for $\epsilon>0$. Since the value at any point of the map (UV) in the neighborhood of the pole is

$$[C\sin(\lambda-\lambda_c)]\cdot[C\cos(\lambda-\lambda_c)]$$

we have

$$C^2\sin(\lambda-\lambda_c)\cos(\lambda-\lambda_c)$$

or

$$\frac{C^2}{2}\sin[2(\lambda-\lambda_c)]\qquad\text{for }\epsilon>0.$$

This pattern, *pattern* 2, is shown in figure 2. In the same manner as above it can be shown that pattern 2 is conservative under addition, subtraction, and the operation of taking the mean.

Maps of the standard deviations of U and of V will have non-negative values everywhere in the neighborhood, $0<\epsilon<\epsilon_M$, but will have indeterminate values at the pole. The standard deviation of V is:

$$S(V)\equiv\sqrt{\frac{1}{N-1}\left\{\Sigma C^2\cos^2(\lambda-\lambda_c)-\frac{1}{N}[\Sigma C\cos(\lambda-\lambda_c)]^2\right\}}.$$

In the neighborhood, not only will $S(U)$ and $S(V)$ be non-negative at every point for $\epsilon>0$, but they will be constant along any meridian since, from above,

$$\frac{\partial}{\partial\phi}[S(V)]=0=\frac{\partial}{\partial\phi}[S(U)]$$

where ϕ=latitude. If the sample is sufficiently large and random, $S(U)$ and $S(V)$ should be almost equal to some positive value throughout the region, $0<\epsilon<\epsilon_M$, and the standard deviations of the quantities X may be taken as positive constants in the neighborhood of the pole.

Maps of the correlation coefficients

$$r(X,Y)\equiv\frac{\text{cov}(X,Y)}{S(X)\cdot S(Y)}\quad\text{and}\quad r(U,V)\equiv\frac{\text{cov}(U,V)}{S(U)\cdot S(V)}$$

will have their zero isopleths in the same place as the corresponding covariance maps, but in general, will not be pattern 1 or pattern 2 maps. Again we should repeat that the functions of U and V are not defined at the poles.

From both physical and mathematical reasonings it can be shown that this method is not applicable for determining the pattern of the divergence fields, such as $\nabla\cdot Q$, in a small neighborhood of the pole.

From the preceding we have formulated the following rules for aid in analyzing maps.

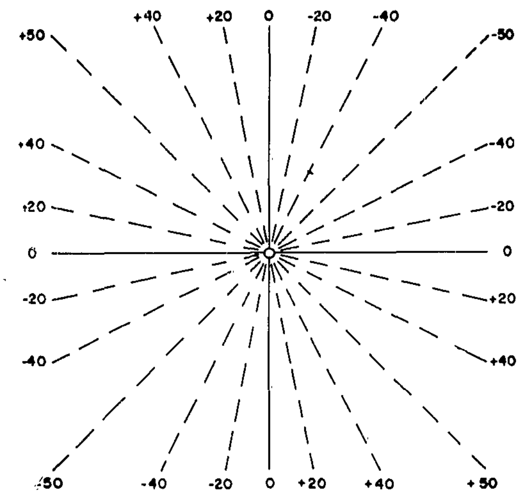

FIGURE 2.—A pattern 2 map, $(C^2/2)\sin[2(\lambda-\lambda_c)]$, for C=10 kt.

3. RULES FOR ANALYSIS OF MAPS OF ALMOST-CONTINUOUS [2] QUANTITIES

1. Zero isopleths of U and V can be drawn as though they pass smoothly through the pole, but the actual values at the pole are not defined.

2. Non-zero isopleths of U and V at the pole are kinked as shown in figure 1.

3. Maps of U and V for the same period have one zero isopleth of U perpendicular to one zero isopleth of V at the pole.

4. The absolute value of any U or V kinked isopleth at the pole can not exceed the value of the wind speed at the pole.

5. At the pole the value of the maximum kinked isopleth is the same for both the U and the V maps.

The following two rules hold anywhere on the maps considered.

6. Any product map (XY) will have zero isopleths where, and only where, the map of X has zero isopleths or the map of Y has zero isopleths, if any exist.

7. A map of the correlation coefficient of X and Y will have its zero isopleths in the same place as the zero isopleths on the map of the covariance of X and Y, if any exist.

[2] Almost-continuous means that there are a countable number of discontinuous points in the field.

The remaining four rules apply only in the neighborhoods of the pole, assuming that $C \neq 0$.

8. In the vicinity of a pole there is one zero isopleth of the map of (XU) which is perpendicular to some one zero isopleth of the map of (XV) at the same pole. If X is non-zero and continuous in the neighborhood of the pole, then these two zero isopleths will be the only zero isopleths, and both the map of (XU) and the map of (XV) will be pattern 1 maps in the neighborhood of the pole.

9. A map of (UV) which includes a pole has perpendicular zero isopleths as shown in figure 2.

10. A map of the covariance of U and V has perpendicular zero isopleths and is a pattern 2 map in the vicinity of the pole.

11. A map of the correlation coefficient of U and V has perpendicular zero isopleths at the pole but is not necessarily a pattern 2 map in the vicinity of the pole.

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Särtryck ur Tellus nr 2, 1960

A Preliminary Study of the Potential to Kinetic Energy Conversion Process in the Stratosphere

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(Manuscript received September 19, 1959)

Abstract

The potential to kinetic energy conversion process in the lower stratosphere associated with the vertical exchange of warm and cold air is evaluated using adiabatically derived vertical velocities for the North American region for a five-day period. Preliminary results suggest the possibility that on the average the kinetic energy of stratospheric motions may not result from a conversion of potential energy within the stratosphere by this process. The further implication is that stratospheric motions are maintained by the motions in the adjacent layers of the atmosphere.

1. Introduction

A fundamental question about stratospheric motions refers to the manner of the maintenance of their kinetic energy. It seems clear that the kinetic energy of the motions of the troposphere is maintained by the conversion from available potential energy as described and verified by many investigators, most recently by LORENZ (1955) and WHITE and SALTZMAN (1956), and that this process is associated with the rising of warm and sinking of cold air. On the other hand, it has not been established that a similar process operates in the stratosphere. This preliminary study represents an attempt to evaluate from observations the nature of this potential to kinetic conversion process in the stratosphere.

2. Procedure

Following WHITE and SALTZMAN (1956) we may write the equations expressing the time rate of change of the kinetic energy of the horizontal wind and the total potential and internal energy of the entire mass of the atmosphere as follows:

$$\frac{\partial}{\partial t}\int_M k\,dm = -\int_M \omega\alpha\,dm - \int_M D\,dm \qquad (1)$$

$$\frac{\partial}{\partial t}\int_M (\Phi + I)\,dm = \int_M \omega\alpha\,dm + \int_M \frac{\partial Q}{\partial t} \qquad (2)$$

where dm is the element of mass and the integration is carried out over the entire mass of the atmosphere, k is the kinetic energy of the horizontal wind, $\omega = \frac{dp}{dt}$ is the individual time rate of change of pressure, α is the specific volume, D is the rate of frictional dissipation of kinetic energy, Φ is the geopotential, I is the internal energy, and $\frac{dQ}{dt}$ is the net rate of heat addition. The appearance of the integral

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$\int_M \omega\alpha dm$ with opposite signs in both equations represents the process of potential to kinetic energy conversion in the atmosphere which is familiarly associated with the vertical exchange of warm and cold air.

The critical problem in evaluating this integral lies in the determination of ω which is closely related to the field of vertical velocity. In this investigation ω was evaluated by the adiabatic relation

$$\omega = -\frac{\dfrac{\partial T}{\partial t} + \mathbf{v}\cdot\nabla T}{\dfrac{\partial T}{\partial p} - \dfrac{\alpha}{c_p}} \tag{3}$$

where T is the temperature, $\mathbf{v}$ is the horizontal wind vector on an isobaric surface, and C_p is the specific heat at constant pressure.

The finite difference form of equation (3) was evaluated from constant pressure charts at 200, 100, 50 and 25 mb over the region of North America extending from 30° to 60°N and from 70° to 120°W. These charts had been carefully analyzed for other purposes by Mr. S. Muench of the Atmospheric Circulations Laboratory, Air Force Cambridge Research Center, and kindly provided us by Mr. W. Hering and Mr. S. Muench for this investigation. The available data consisted of seven 200, 100, and 50 mb and six 25 mb charts covering the period 28 January to 3 February, 1957, a period of intense stratospheric warming as descibed by CRAIG and HERING (1959). The wind velocities were evaluated geostrophically and the temperatures hydrostatically. Contour height values were abstracted at a grid system of 273 points spaced 2.5 degrees apart and the ω's were obtained at a grid system of 60 points spaced 5 degrees apart. The time derivatives were approximated by finite differences over 24 hour intervals. The computation yielded fields of the 24-hour average values of ω and α for the layers 200—50 mb and 100—25 mb which were taken to be representative of these fields at the 100 and 50 mb levels.

Using the values of ω and α thus derived it was possible to sample the integrand for this region of the hemisphere for the short period of time as indicated, and for the layers of the atmosphere at approximately the height of the 100 and 50 mb levels. Average values of the integrand $\omega\alpha$ over space and time were computed and an analysis of the covariance of $\omega\alpha$ was performed in a manner entirely analogous to that of WHITE and SALTZMAN (1956). Following their procedure we may write:

$$\overline{\{[\omega\alpha]\}} - \overline{\{[\omega]\}}\,\overline{\{[\alpha]\}} = \overline{\{[\omega]''[\alpha]''\}} + \overline{\{[\omega'\alpha']\}} + \overline{\{[\omega]\}'''\{[\alpha]\}'''} \tag{4}$$

where the single, double and triple primes denote deviations from east-west, north-south, and time averages respectively, the brackets indicate an average in the east-west direction, the braces an average in the north-south direction and the bars an average in time.

The left side of equation (4) is a more representative measure of the integrand average than the term $\{[\omega\alpha]\}$ alone, since the second term on the left side must vanish when the average is taken over the entire hemisphere and will give rise to spurious non-zero values if not subtracted out because of the limited area treated. The first term on the right is a measure of the conversion process due to overturnings in the north-south direction, the second is a measure of the process due to overturnings in the east-west direction. The last term is associated with temporal pulsations of the space average values of ω and α.

3. Results

The values of ω obtained are reasonable. When transformed by means of the hydrostatic equation to equivalent vertical velocities, the standard deviation is 0.86 cm sec^{-1} at 100 mb and 1.35 cm sec^{-1} at 50 mb. A typical example of the distribution of the vertical velocity in relation to the temperature and contour height fields at 50 mb is shown in Fig. 1 for 29 January, 1957.

The numerical values of the potential to kinetic energy conversion processes are shown in Table 1.

Table 1. The rate of conversion between potential and kinetic energy. A minus sign indicates a conversion from potential to kinetic energy. Units in ergs gm^{-1} sec^{-1}.

	1	2	3	4
	$\overline{\{[\omega\alpha]\}} - \overline{\{[\omega]\}}\,\overline{\{[\alpha]\}}$	$\overline{\{[\omega]''[\alpha]''\}}$	$\overline{\{[\omega'\alpha']\}}$	$\overline{\{[\omega]\}'''\{[\alpha]\}'''}$
50 mb	7.5	—0.1	7.5	0.1
100 mb	—0.5	—0.3	—0.4	0.2

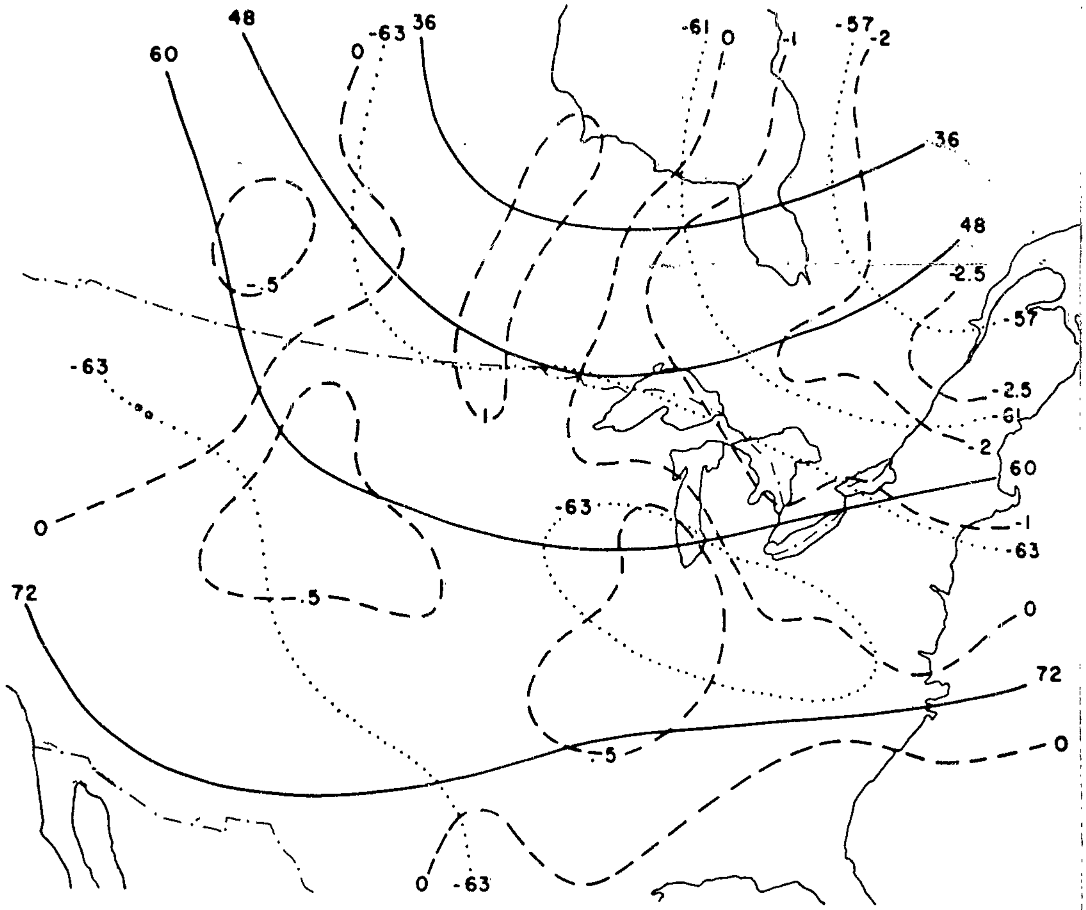

Fig. 1. The vertical velocity, temperature, and contour height distribution at 50 mb for January 29, 1957, 03 GMT. Vertical velocity units in cm sec^{-1}.

The first column of Table 1 indicates that at 50 mb the conversion process is opposite to that found in the troposphere being from kinetic to potential energy in this region of the stratosphere. The sense of this conversion process was the same on each of the four individual days examined. At 100 mb the magnitude of the conversion term is not essentially different from zero and the six individual daily values on which the mean is based vary in sign.

A more detailed analysis of this conversion process is given in columns 2, 3, and 4 of table 1. At 50 mb the principal contribution to the conversion process is associated with overturnings in the east-west direction as indicated in column 3. The contributions of the other two terms in columns 2 and 4 are negligible. At 100 mb the conversion rates associated with each of the individual terms are small.

Since the 100 mb data are representative of the layer between 200 and 50 mb, this layer will be wholly contained in the stratosphere only at high latitudes and will be partially within the troposphere at low latitudes. It should be possible upon examination of the latitudinal variation of the term $[\overline{\omega'\alpha'}]$ to detect whether

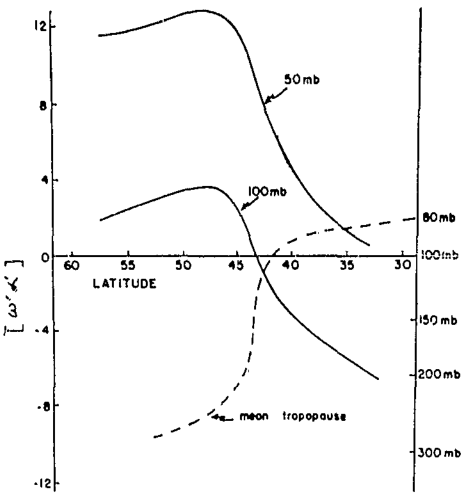

Fig. 2. The latitudinal variation of the potential to kinetic energy conversion process due to east-west overturnings in units of ergs gm^{-1} sec^{-1}. Also shown is the latitudinal variation of the mean tropopause height as a function of pressure.

there is any systematic difference in the sense of the conversion process as one proceeds from troposphere to stratosphere. The latitudinal variation of this term at both 100 and 50 mb together with a plot of the mean tropopause height is shown in Fig. 2.

It can be seen that the sign of the conversion process changes at approximately that latitude (43°N) where the mean tropopause crosses the 100 mb level, being from kinetic to potential to the north and from potential to kinetic to the south of this latitude. This observational feature agrees with the concept of a reversal of the potential to kinetic energy conversion process from troposphere to stratosphere.

If these observational findings are further substantiated by more extensive investigations on a hemispherical basis and over longer periods of time then the implications seem far-reaching. If the kinetic energy of the stratospheric motions is not maintained by conversion from potential energy within the stratosphere by the processes associated with the vertical exchange of warm and cold air then it must be maintained by boundary interactions with the adjacent layers above and below. Such processes involve the vertical transport of existing kinetic energy through the top and bottom boundaries of the layer as pointed out by STARR (1959) and other boundary processes associated with variations in the height of the bounding pressure surfaces. Should general verification of these concepts be obtained, one possible implication would be that stratospheric motions must to a large extent vary in response to tropospheric changes, and that the explanation for many observed characteristics of the stratosphere, such as the size and motion of circulation systems and the seasonal and latitudinal distribution of ozone, may very well lie in a better understanding of the linkages between the stratosphere and troposphere.

4. Critical Remarks

As in all such limited investigations care must be exercised in generalization before confirmation on the basis of more extensive data. This particular study suffers from the following deficiencies:

a. The results are based on a sample of data from a small area of the hemisphere and for a very short period of time.
b. The period of time studied was one of abnormal high level temperature changes, and the results may not be typical of more normal conditions.
c. The vertical velocity computations on which the results fundamentally depend are based upon the adiabatic assumption.

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PRINTED IN SWEDEN
ESSELTE, STHLM 60

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Report No. AFCRL - 63 - 435

STUDIES OF THE STRATOSPHERIC GENERAL CIRCULATION
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Unclassified

1. General Circulation of the Atmosphere
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